

COMPLEX MONITORING OF VOLCANIC ACTIVITY

METHODS
AND RESULTS

Vyacheslav M. Zobin
Editor

Earth Sciences in the 21st Century

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COMPLEX MONITORING OF VOLCANIC ACTIVITY METHODS AND RESULTS

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VYACHESLAV M. ZOBIN
EDITOR



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PREFACE

This book presents the collection of papers devoted to monitoring volcanic activity with seismic, acoustic, geodetic, and video observations, and the catalogs of volcanic eruptions. The volcanoes, representing the subject of these observations (Bezymianny, Etna, Karymsky, Kizimen, Sakurajima, Sheveluch, Stromboli, Popocatépetl, Puracé, Volcán de Colima), are between the most active on our Earth.

Three first Chapters are devoted to different aspects of quantification of Vulcanian activity of volcanoes: the quantification of Vulcanian explosions from the seismic and acoustic records (Chapters 1 and 2), and the estimation of the height of eruption plumes of Vulcanian explosions from the seismic records (Chapter 3).

Next three Chapters are concerned with monitoring volcanic eruptions and their forecasting. They are based on the changes in the spectral characteristics of volcanic tremor (Chapter 4), on the statistical analysis of appearance of Strombolian explosions (Chapter 5), and on the analysis of historical eruption time series (Chapter 6).

Next two Chapters are devoted to monitoring of the deformation of the volcanic edifice of Volcán de Colima and the building of lava dome in the crater of this volcano. Chapter 7 discusses the resolution of different-scale geodetic observations at Volcán de Colima, using remote sensing and ground-based techniques, during monitoring of short-term volcanic events, such as Vulcanian explosions, and long-term volcanic events, such as the effusive and explosive stages of eruptions and the general tendencies in the development of the volcanic structure. Chapter 8 illustrates the application of continuous video observations for monitoring the 2007-2012 lava dome growth in the crater of volcano.

Chapter 9 presents a review of the seismic activity associated with the geothermal appearance of volcanic activity.

I am very grateful to the scientists who reviewed the manuscripts submitted to this book. They did a hard work not only giving their valuable comments but also correcting English grammar of the most of texts. Among them are Alejandra Arciniega (Universidad Nacional Autonoma de México, México), Andrea Cannata (Istituto Nazionale di Geofisica e Vulcanologia, Italy), Salvatore Gambino (Istituto Nazionale di Geofisica e Vulcanologia, Italy), Sharon Kedar (Jet Propulsion Laboratory, U.S.A.), Kostas Konstantinou (National Central University, Taiwan), John Londoño (INGEOMINAS, Colombia), John Lyons (Escuela Politécnica Nacional, Ecuador), Adriano Mazzini (University of Oslo, Norway), McNutt (University of Alaska, U.S.A.), Mauricio Mora (Universidad de Costa Rica, Costa Rica), Edoardo del Pezzo (Istituto Nazionale di Geofisica e Vulcanologia, Italy), Virginie Pinel (Université de Savoie, France), Graham Ryan (University of Auckland, New Zealand),

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I thank *Nova Science Publishers Inc.* for their kind invitation to prepare this book. It was a pleasure to collaborate with the colleagues-volcanologists submitting the results of their studies to our book. Most of them, as well as the volcanoes described in their chapters, are my good friends.



Vyacheslav M. Zobin
Colima, México, 20 August 2012

Chapter 1

QUANTIFICATION OF THE 1999 VULCANIAN EXPLOSIONS AT SAKURAJIMA VOLCANO, JAPAN FROM BROADBAND SEISMIC RECORDS

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ABSTRACT

Fourteen Vulcanian explosions that occurred at Sakurajima volcano, Japan in 1999 were quantified in Joules using methodology of fast determination of the energy of explosions derived from broadband seismic records. The comparison of the energy of these explosions with the seismic moments estimated for the same events showed that these two size parameters are good correlated with the coefficient of correlation of 0.94.

1. INTRODUCTION

The goal of quantifying the size of explosive volcanic eruptions has produced several magnitude and intensity scales, including the eight-grade Volcanic Explosivity Index, or VEI (Newhall and Self, 1982; Pyle, 2000). These scales are based primarily on the size of volcanic deposit and, as a result, their principal subdivisions cover large ranges of values (*e.g.*, ranges in volume by about a factor of 10 for each VEI unit). In addition, they are commonly applied to the total products of an eruption, as opposed to those of specific phases during a related sequence of explosive events. An alternative approach to quantifying size is to use the seismic signals induced by volcanic explosions.

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Although such an approach is necessarily restricted to eruptions that are being monitored with broadband instruments, it has the potential advantage of discriminating the magnitudes of events with greater precision (Kanamori et al., 1984; Nishimura, 1995; Chouet et al., 1997, 2008; Tameguri et al., 2002; Zobin et al., 2006; 2009).

Vulcanian explosions, characteristic for andesitic and dacitic volcanoes, are a good subject for this type of quantification. They are short (seconds to minutes) discrete explosions characterized by a small amount (less than 1 km^3) of erupted products and high eruption columns, which can reach 10-20 km altitude, and occupy the intermediate position in the explosion hierarchy between Strombolian explosions producing eruption columns of tens or hundreds of meters in height and Plinian explosions whose eruption columns may reach up to 45 km altitude (Francis, 1993). Their discrete character allows having a good quality broadband records of each of them enough separated in time.

The quantification of volcanic explosions from broadband seismic records has been developed in two directions. The first of them is based on the representation of the source of an explosion as a centroid (e. g., Chouet et al., 2008) that allows estimating source mechanisms of explosions from moment-tensor inversions of long-period data of the total seismic record, band-pass filtering broadband seismic records in the 2-30 Hz band.

The second approach for source mechanisms of explosions (Tameguri et al., 2002; Zobin et al., 2006; 2009) is based on the detailed study of the sequence of individual phases of the seismic records associated with volcanic explosion, considering an explosion as a multi-staged process. Tameguri et al. (2002) discriminated four phases of the seismic records of Vulcanian explosions at andesitic Sakurajima volcano and estimated moment tensors for the sources of these phases. Zobin et al (2006; 2009) developed a two-stage model of explosive process which allows quantifying the volcanic explosions in the units of force (Newtons) and energy (Joules) from unfiltered broadband seismic records. This method was successfully applied for the quantification of Vulcanian explosions produced by two Mexican andesitic volcanoes, Popocatepetl and Volcán de Colima (Zobin et al., 2009; Zobin and Martinez, 2010).

In this Chapter, we analyze broadband seismic records associated with fourteen 1999 Vulcanian explosions of Sakurajima volcano. We discuss the moment tensors calculated by Tameguri et al. (2002) for different stages of explosive process and apply the methodology, proposed by Zobin et al. (2009), to quantify the energy of these events.

2. SAKURAJIMA VOLCANO AND ITS EXPLOSION ACTIVITY

Sakurajima is an andesitic stratovolcano situated in South Kyushu, Japan, and reaching 1,117 m above sea level (a.s.l.) (Figure 1). The volcano has three craters aligned in NS direction and forming a ridge surrounded by steep slopes. Eruptive activity at the summit cone ended about 4900 years ago, after which eruptions took place at the southern-most cone Minamidake (1,040 m a.s.l.).

In the historical time, large eruptions were originated three times in 1471-1476, 1779-1781, and 1914. Each eruption ejected lava and pyroclastic materials of 1 to 2 km^3 from the both flanks of the volcano. The current eruptive activity, summit eruption, started in October 1955, and has continued up to the recent date (BGVN, 2000). The Sakurajima Volcano

Research Center, which carries out the monitoring of the Sakurajima activity, is a good equipped observatory with the different type of instruments for study of Vulcanian explosions.

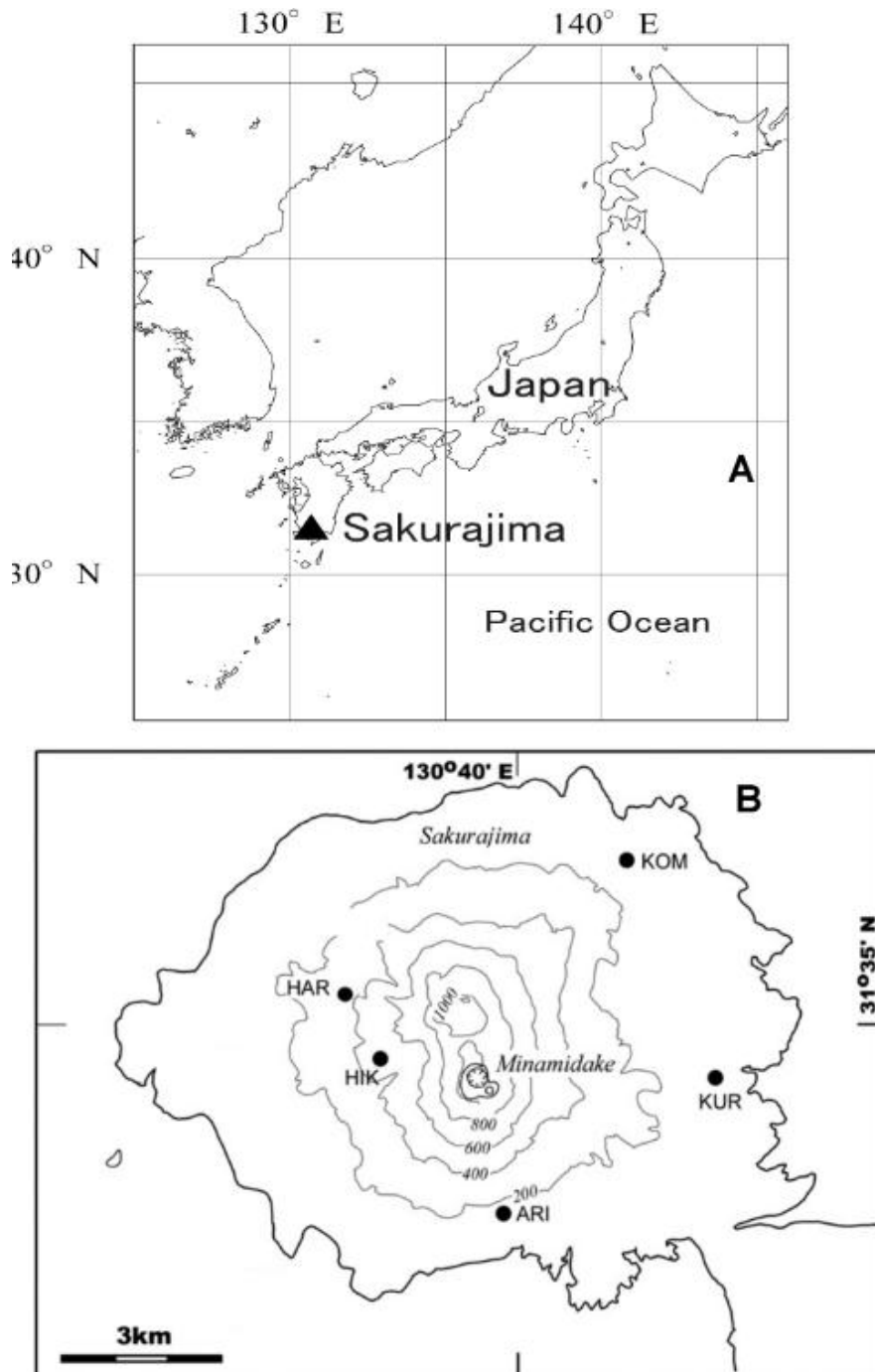


Figure 1. Position of Sakurajima volcano (A) and the location of broadband seismic stations shown as the filled circles in (B). The contour lines at 200, 400, 600, 800, and 1000 m show the general relief.

The hypocenters of explosion earthquakes at Sakurajima were located filling the cylindrical-shape zone at depths of 1 to 3 km beneath the summit crater (Iguchi, 1994). Tilt and strain steps, which were observed at the onset of explosive eruptions, allowed to Ishihara (1990) to conclude that these steps were caused by an instantaneous pressure decrease that reflects outbreak of the gas pocket formed by storage of volcanic gases at the uppermost part of volcanic conduit. Uhira and Takeo (1994) obtained from the inversion of long-period records of two 1986 explosive earthquakes that the moment tensor diagonal components correspond to a buried vertical cylindrical source ($M_{xx}:M_{yy}:M_{zz} = 2:2:1$).

The contribution of the vertical single force component calculated for these events was small comparing to moment tensor components. Uhira and Takeo (1994) concluded that the 1998 Sakurajima explosions were produced by two sources situated at the different depths. They postulated that the deep source was a region of magma intrusion and the shallow source was the source of explosions.

2.1. The 1999 Explosive Activity at Sakurajima Volcano

A sharp increase in explosive activity at Sakurajima volcano was observed during July 1999 to February 2000 with a peak activity during December 1999 (Figure 2).

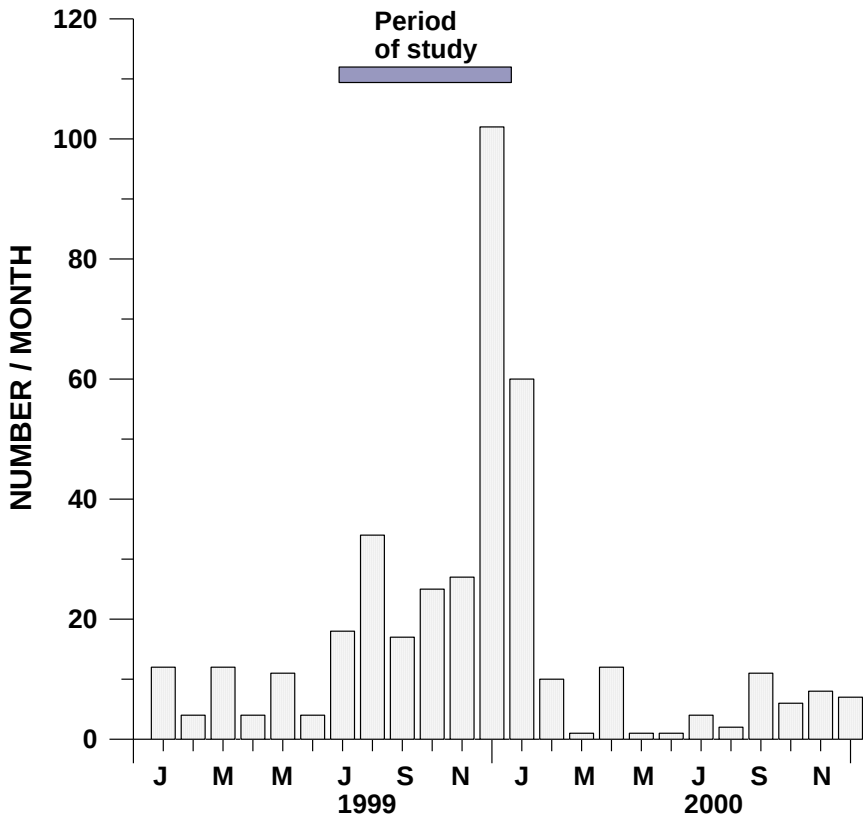


Figure 2. Monthly number of explosions at Sakurajima volcano during 1999 and 2000 (BGVN, 2000).

Vigorous activity during December was marked by incandescent columns and large amounts of bomb ejections.

According to a JMA field inspection, ballistics were scattered 3-4 km away from the Minamidake crater; the maximum size was 4 cm across. Daily numbers of eruptions ranged from 2 to 8 during early- to mid-December; eruptions were mostly explosive. The plume heights of the December explosions ranged from 1,500 m to 2,000 m. The total of 88 explosions took place on 23 consecutive days between 3 and 25 December. This was the longest record of daily explosions since JMA started observing Sakurajima in 1955 (BGVN, 2000). The detailed study of the 1999 Sakurajima explosions was performed by Tameguri et al. (2002). Temporary broadband seismic observations around Sakurajima volcano were carried out during the period from August to December 1999 (Figure 1b).

Five seismic stations were installed at the distances of 2.5 km to 4.7 km from the crater Minamidake. The instruments had a flat velocity response from 0.02 to 120 s.

2.3. The Seismic Phases Associated with the Sakurajima Explosions

The fine structure of broadband seismic records of Vulcanian explosions at Sakurajima volcano occurring during August-December 1999 was discussed by Tameguri et al. (2002). Figure 3 shows the typical broadband seismic records of the 1999 Sakurajima explosion obtained on three seismic stations situated at the distances of 2.7, 4.1, and 4.7 km from the crater. It is possible to distinguish three groups of the seismic signals clearly seen on all seismograms: the first group of low-amplitude short-period pulses *SP1* (between the arrivals 1 and 2), the group of high-amplitude long-period pulses *LP* (between the arrivals 2 and 3) and the second group of high-amplitude short-period pulses *SP2* (between the arrivals 3 and 4). The phases composing the initial short-period pulses were identified by Tameguri et al. (2002) as *P*-waves, and the long-period *LP* phase, as composed of Rayleigh waves.

Moment tensors of the seismic sources, that generated the phases *SP1* and *LP*, were estimated from the waveform inversions. The sources of the waves composing the short-period initial waveforms, were characterized by three diagonal components, corresponding to the isotropic expansion source, and by the horizontal dipoles twice large than the vertical, corresponding to a volume change of cylindrical source (2:2:1), respectively. These sources of short-period vibrations were located at depths of 2 km beneath the crater.

The *LP* phase was proposed to be excited by the third source, characterized by an isotropic expansion and the following horizontal contraction, and located at shallow depths of 0.25 to 0.5 km beneath the crater bottom.

Tameguri et al. (2002) found that the origin times of the *LP* phases coincided with those of the air-waves associated with the explosions. More recent studies carried out by Yokoo et al. (2009) showed that the origin times of the *LP* phases of the Sakurajima explosions coincided with the preceding low-amplitude phase of the infrasonic wave. Yokoo et al. (2009) showed that the origin of the impulsive compression phase of the infrasonic wave coincided with the peak time of the moment velocity reconstructed for the *LP* phase. Therefore, the explosion occurs after the beginning of the radiation of the *LP* phase. Based on the conceptual model of the Sakurajima explosions, proposed by Ishihara (1990), the sequence of the seismic sources may correspond to the following eruptive process.

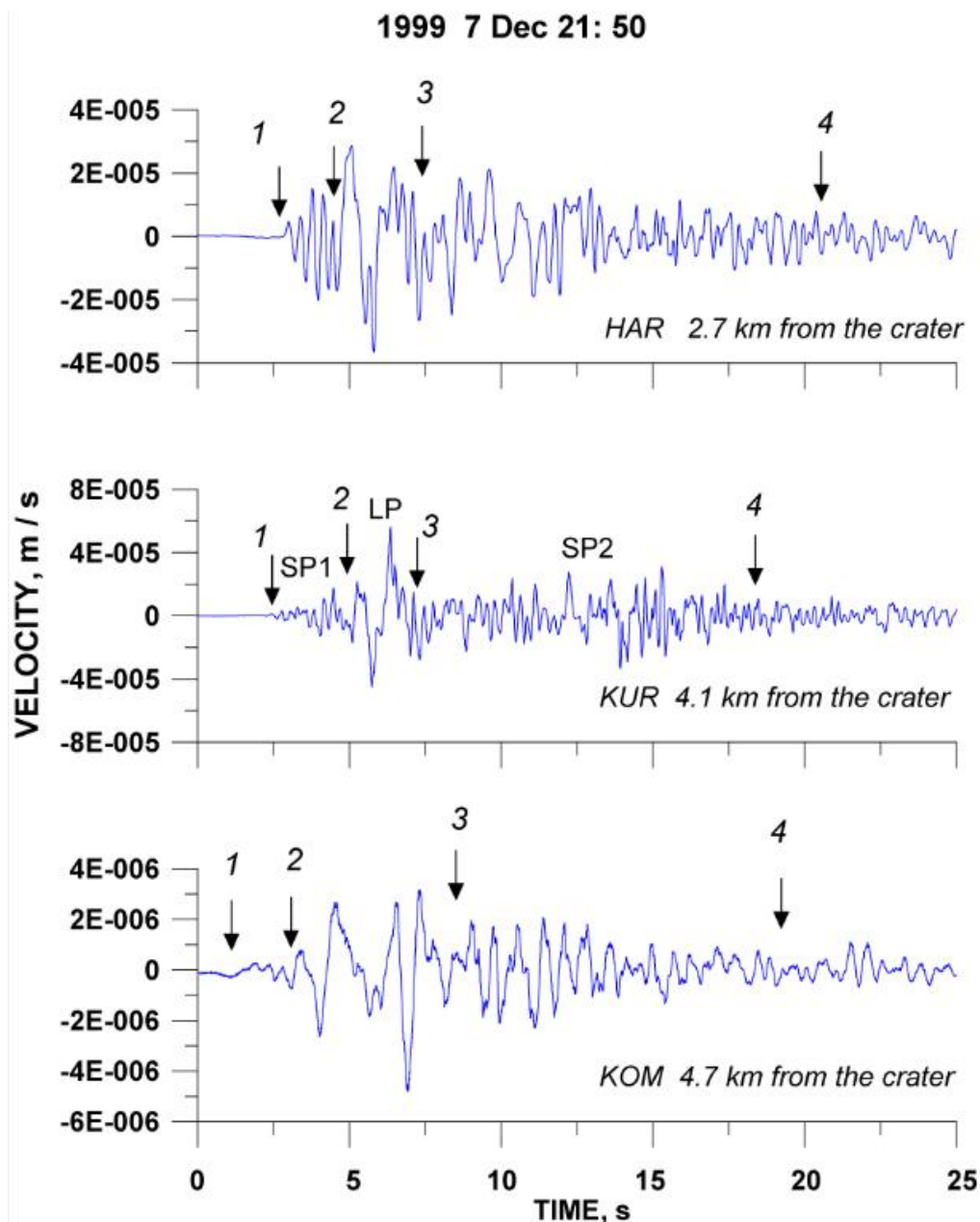


Figure 3. The seismic velocity signals (vertical component) associated with the Sakurajima Vulcanian explosion of 7 December 1999, 21:50. HAR, KUR, and KOM are the codes of seismic stations. 1, 2, 3, and 4 are the arrivals of the different seismic signals, see in the text.

The gas pocket, created by the ex-solution of volatiles in the upper part of the shallow magma chamber at the depth about 2 km (short-period phases), abruptly moves up through the conduit to the surface generating the phase *LP*.

Then the explosion, accompanying by the outflow of gas, ash and volcanic blocks, occurs (the coda of *LP* signal and the second high-frequency group of the seismic record).

3. QUANTIFICATION OF THE 1999 EXPLOSIONS AT SAKURAJIMA VOLCANO USING ZOBIN ET AL'S METHODOLOGY

Fourteen explosion earthquakes (Table 1), recorded by the temporary network from 27 August to 24 December 1999 and quantified by Tameguri et al. (2002) in the terms of seismic moment, were then quantified in the terms of energy applying the methodology described in (Zobin et al., 2006; 2009).

3.1. Two-Stage Conceptual Model of the Explosive Event

The two-stage conceptual model of explosive event was developed in Zobin et al. (2006; 2009). This conceptual model of Vulcanian explosions describes the volcanic explosion process as consisting of the movement of the fragmented magma to the surface (first stage) and a subsequent explosion (second stage). The degassing of silicic magmas is proposed to occur by gas filtration through a system of interconnected bubbles. It is proposed that the modeling of long-period group of seismic pulses, shown in Figure 3 between arrows 2 and 3, may give the size of force governing the vertical movement of magma, and the subsequent high-frequency pulse *SP2* (shown between arrows 3 and 4), associated with the explosion and the outflow of volcanic products, may be used for estimation of the energy of explosion.

Table 1. Data about the 1999 Sakurajima explosions

Date, mmdd hhmm		Energy, J	M_{zz} , N-m
0827	1921	1.08E+08	2.70E+11
0903	455	6.72E+07	2.80E+11
0915	414	3.12E+07	2.40E+11
1006	127	3.04E+07	2.50E+11
1116	1106	3.74E+07	2.90E+11
1206	1813	2.17E+08	5.50E+11
1207	2054	4.72E+08	7.60E+11
1207	2150	9.50E+08	9.90E+11
1210	1903	4.87E+09	1.18E+12
1212	1648	4.03E+08	1.07E+12
1217	1530	1.01E+09	1.01E+12
1220	732	4.30E+08	6.70E+11
1221	1317	2.59E+08	4.70E+11
1224	554	1.53E+09	1.28E+12

Note. M_{zz} values were taken from Table 4 of Tameguri et al. (2002).

These two magnitudes are good correlated (Zobin et al., 2009); therefore, it is enough to estimate only one of them, the energy, for fast explosion quantification.

3.2. Estimation of the Explosion Energy

The group of high-amplitude short-period pulses *SP2* (between the arrows 3 and 4 in Figure 3), which are considered as associated with the explosion and the outflow of volcanic products, was used to estimate the energy of explosions. The theorem of Parseval states that it is possible to calculate the total amount of energy E_i (for the unit of mass) in a continuous signal v (velocity) from the Fourier spectrum of the signal:

$$E_i = \int_0^{2\pi} v^2(t) dt \quad (1)$$

where t is a time of seismic vibration. According to this theorem, the energy E_i (in Joules) of the seismic pulse recorded at i -station may be calculated as the integral (1) where v is the length of the Fourier vector taken from the minimum to maximum frequencies of the spectrum of the seismic velocity record. To estimate the total energy of the explosion, one can multiply the energy of the *SP2* pulse recorded at i -station, E_i , by the coefficient of the seismic portion of the total energy of the explosion (the ratio between seismic and total energy) p and the coefficient of effective attenuation of seismic energy with distance including the effects of geometrical spreading, energy absorption, focal mechanism and station conditions, k . The coefficients p and k were approximately estimated for the Volcán de Colima explosions recorded by broadband seismic station at a distance of 4 km from the crater as $p = 10^{-5}$ and $k = 10^{-13}$ (Zobin et al., 2006). It allowed us calculating the energy of an explosion from the equation

$$E = 10^{18} \times E_{4km} \quad (2)$$

Equation (2) provides the values of E with a precision of about half an order of magnitude. At the same time, the estimations of E for a sequence of explosions at the same volcano can give a resolution of about ± 0.1 of log unit.

3.3. Processing of the Seismic Records

Figure 4 shows the processing of the seismic record to obtain the energy of the seismic pulse at i -station. The *SP2* pulse, following the long-period waveform, was taken at the vertical component of the seismic velocity signal (Figure 4A). For this selected signal, the Fourier spectrum was calculated (Figure 4B). Then the energy of this *SP2* pulse was estimated according to equation (1).

We used the seismic records of three broadband seismic stations, HAR, KUR and KOM, situated at the distances of 2.7, 4.1 and 4.7 km from Minamidake crater, respectively. The

seismograms of station HIK were not accessible; the seismic records of station ARI demonstrated a strong site effect and were not used for energy determination.

The distance of 4.1 km from the crater of station KUR is the same as the distance of broadband seismic station at Volcán de Colima that was used for obtaining of equation (2). It allows us to re-write the equation (2) as

$$E = 10^{18} \times E_{KUR}. \quad (3)$$

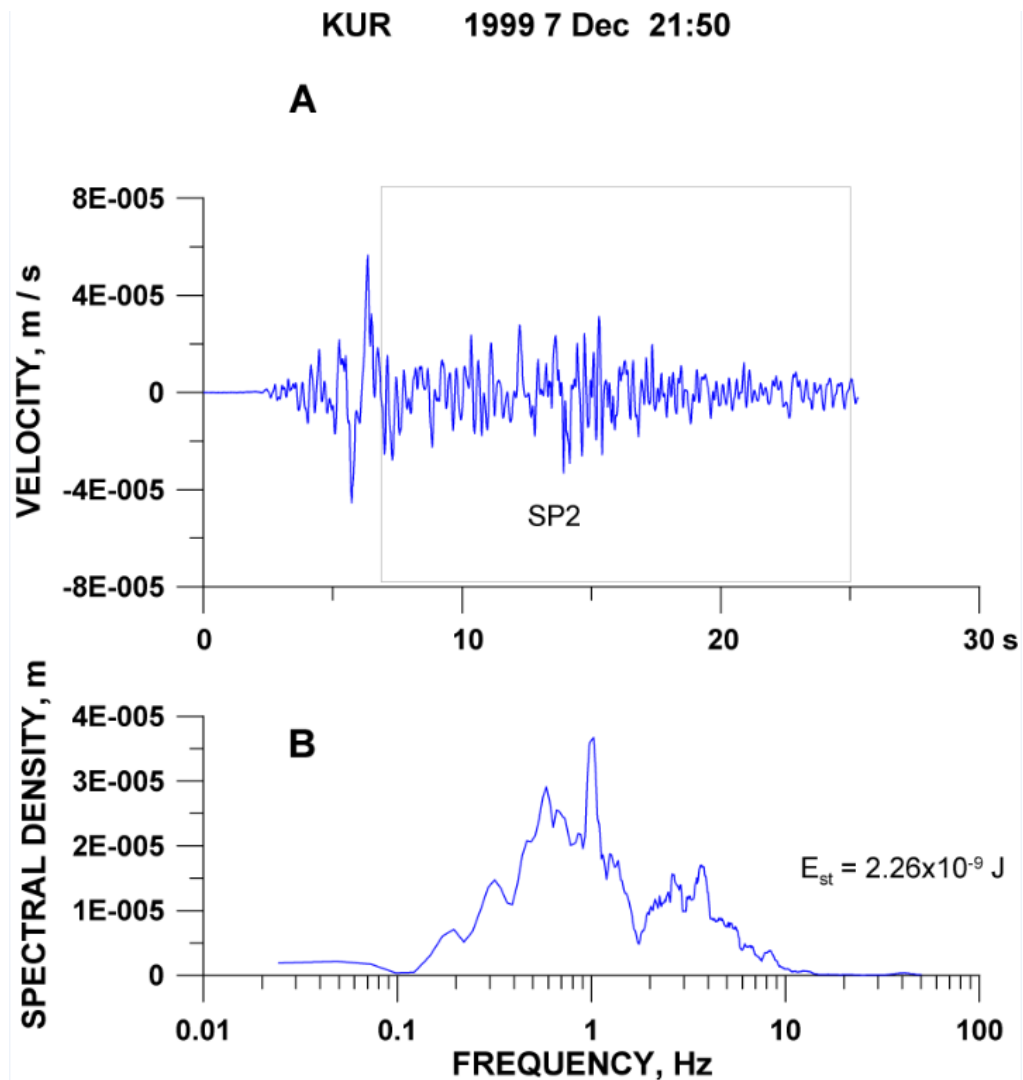


Figure 4. Processing of the unfiltered seismic signal corrected for instrumental response (velocity, vertical component) to obtain the energy of an explosion. It is shown SP2 pulse within the rectangle (A) and its Fourier spectrum (B).

To have possibility to use the seismic records of other stations, the transfer coefficients were estimated. The equation (3) would be for the seismic records of station HAR as

$$E = 0.78 \times 10^{18} \times E_{HAR}$$

(4)

and for station KOM as

$$E = 3.45 \times 10^{18} \times E_{KOM}.$$

(5)

The energy of 14 explosions was calculated as the means of values obtained at three stations.

3.4. Energy Characteristics of the 1999 Sakurajima Explosions

Table 1 shows the energy estimated for fourteen 1999 Sakurajima explosions. They range from 3.0×10^7 J to 4.9×10^9 J.

Five explosions with energy between 3.0×10^7 J and 1.1×10^8 J were recorded during the August-November activity; the energy of the December explosions varies from 2.2×10^8 J to 4.9×10^9 J.

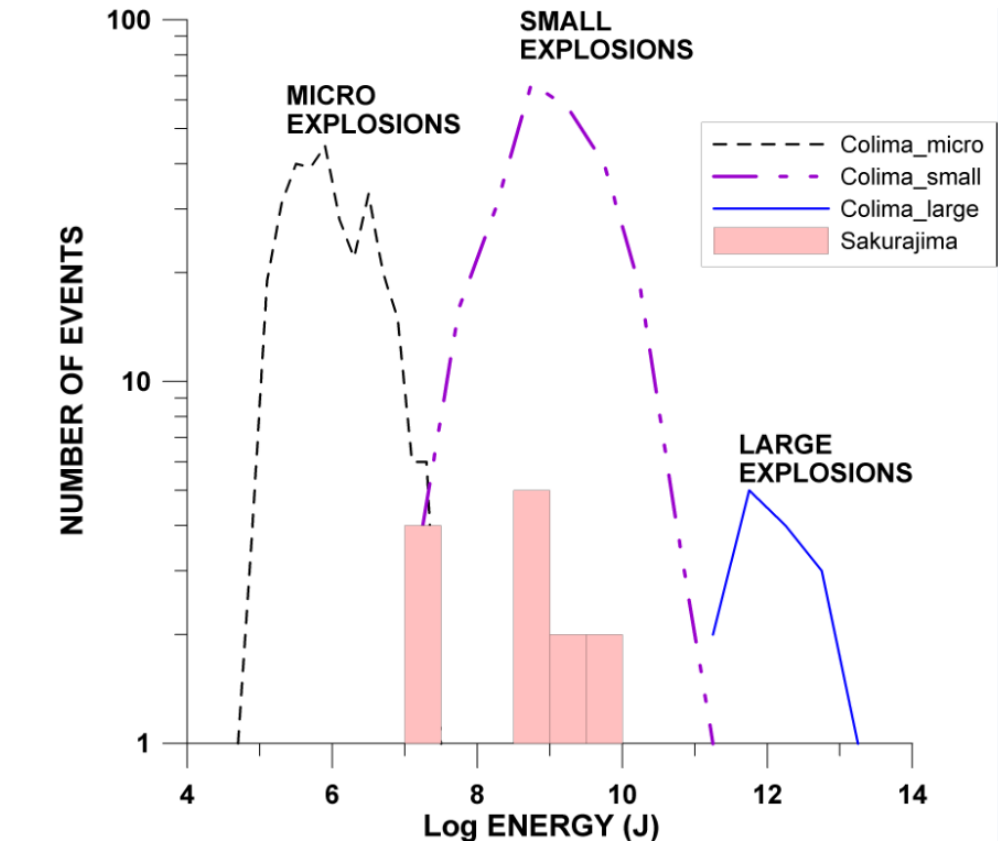


Figure 5. Position of the Sakurajima explosions (columns) within the Vulcanian explosion hierarchy. The distributions of large, small and micro explosions observed at Volcán de Colima (Zobin et al., 2010) are shown demonstrating the hierarchy of Vulcanian explosions.

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Our list of the 1999 explosions is not complete, but for the studied events, the largest energy was obtained for the explosion that occurred on 10 December at 19:03.

The experience with the quantification of the Colima Vulcanian explosions allowed discriminating them into three categories: large, small and micro (Zobin et al., 2010). The separation between large and small explosions is at a level of 10^{11} J and the separation between small and micro-explosions is at a level of 10^8 J.

Figure 5 shows that, according to this classification, four September-November Sakurajima explosions may be considered as micro-explosions, whilst the August event as well as the December 1999 events, as small explosions.

We compare the energy quantification of the 1999 Sakurajima explosions with the quantification of the same events from moment tensor analysis (Tameguri et al., 2002).

The components of moment tensor calculated for *LP* phase describe the magnitude of explosive process. Figure 6 shows the maximum likelihood fitting of the relationship between the components of moment tensor, estimated by the inversion of *LP* phase and the energy of Sakurajima explosions. It was obtained the equation

$$\text{Log } M_{zz} = 0.39 \text{ Log } E + 8.45 \quad (6)$$

The coefficient of correlation between these parameters $R = 0.94$.

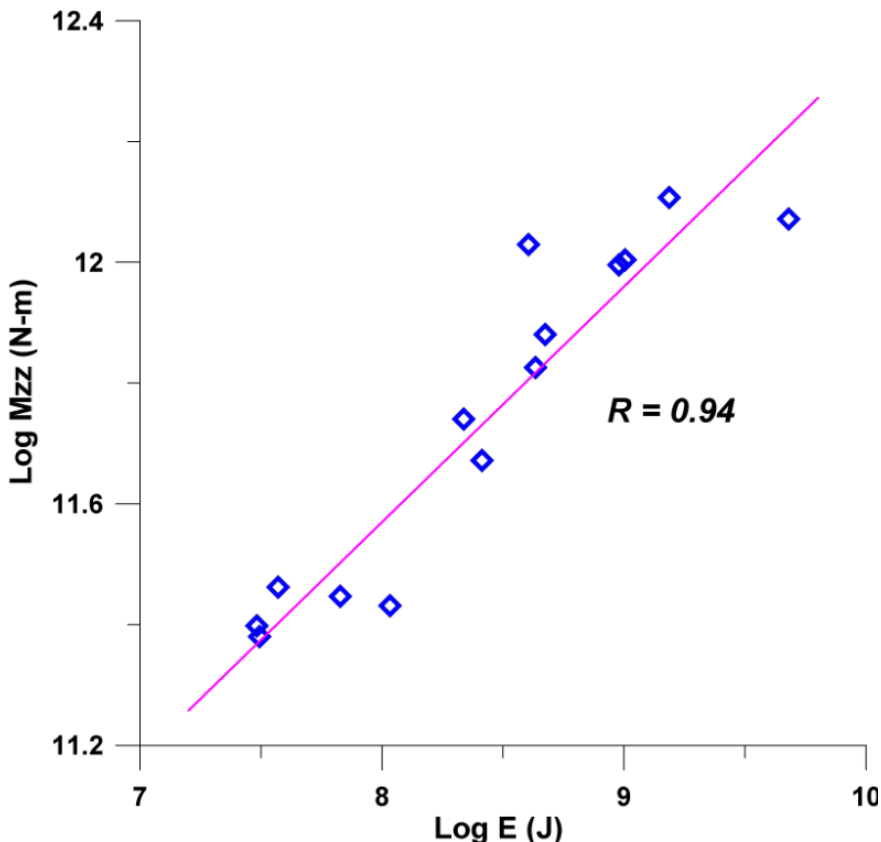


Figure 6. Relationship between the components M_{zz} of moment tensors of Sakurajima explosions estimated in (Tameguri et al., 2002) and the energy of the same events.

CONCLUSION

The application of two methodologies for quantification of the 1999 Sakurajima explosions showed that the components of the seismic moments and the energy of the explosions are good correlated. It demonstrates the possibility of fast quantification of the explosive events using the methodology of energy estimation proposed by Zobin et al. (2006).

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Chapter 2

OBSERVATION OF VULCANIAN EXPLOSIONS WITH SEISMIC AND ACOUSTIC DATA AT POPOCATÉPETL VOLCANO, MEXICO

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ABSTRACT

A series of Vulcanian explosions at Popocatepetl volcano during 2008-2012 was studied, with an examination of the associated acoustic-seismic data to improve our knowledge of the generation of this kind of events. The seismic data indicate a complex nature of simultaneous processes associated with the volcanic explosions. Four distinct frequency bands were analysed: very-long-period (VLP), long-period (LP), short-period (SP), and high-frequency (HF). The VLP seismic signals were recorded only for the largest events with seismic energy greater than 115 MJ at Tlamacas (PBMV) station, and of 1691 MJ at Colibrí (PBCV) station. The seismic energy of the 2008-2012 Popocatepetl explosions was comparable with the seismic energy reported for the majority of Vulcanian explosions at andesitic volcanoes around the world. In the case of acoustic energy, it was slightly higher for Popocatepetl explosions as compared to other volcanoes. The Volcanic Acoustic–Seismic Ratio (VASR) was greater than 1 for the

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majority of the explosions indicating that the acoustic energy of the Popocatepetl explosions was greater than the seismic energy. The delay times between the seismic and the acoustic signals show that the depth of the acoustic-seismic signals sources was shallow, approximately between 0 and 220 m beneath the lava dome or crater floor.

I. INTRODUCTION

Popocatepetl volcano is located in the central zone of the Trans-Mexican Volcanic Belt (TMVB) (Figure 1), and after a 67-year period of quiescence began a new phase of activity with an ash and gas discharge on 21 December 1994. Its previous episode of activity lasted from 1920 to 1927, generating lava domes within its crater, as well as explosions and ash emissions. The current episode of activity has also witnessed the growth of lava domes within the crater, as well as Vulcanian explosions with a maximum VEI of 3. The seismic activity associated with the recent Vulcanian explosions at Popocatepetl has been studied by many authors (Arciniega et al., 1999; Cruz-Atienza et al., 2001; Arciniega et al., 2008; Chouet et al., 2005; Zobin et al., 2008, 2009; 2010). In these studies, some interesting results have been found such as the occurrence of Very-Long-Period events (VLPs) associated with the explosions. To obtain VLP phases, Arciniega et al., (1999) filtered the vertical component of the seismic signals between 25 and 100 sec. The polarization of the VLPs permitted the location of two sources of the explosions: one being located 1 km below the crater and the second source appeared to be deeper. The inversion of VLP signals of Popocatepetl explosions allowed to Chouet et al. (2005) to show that the sources of VLPs associated with Vulcanian explosions were located at a depth of 1.5 km beneath the western perimeter of the summit crater. This source was considered to be composed of a shallow dipping crack intersecting a steeply dipping one, whose surface extension bisects the vent. Zobin and Martínez (2008) observed that the seismic records of the Popocatepetl explosions were composed of a low frequency phase, followed by a phase of greater frequency and amplitude.

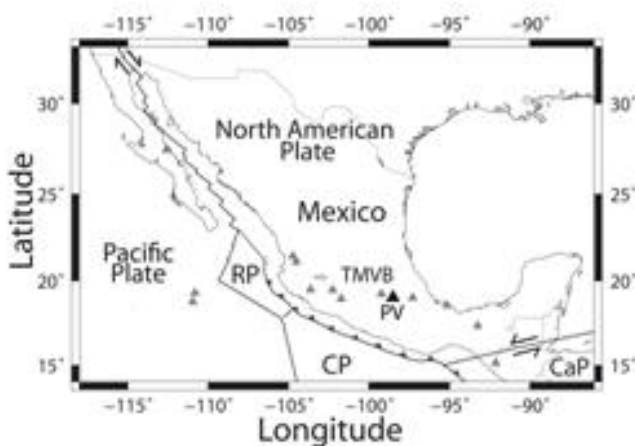


Figure 1. Location of Popocatepetl volcano (PV) inside the Trans-Mexican Volcanic Belt (TMVB). CP means Cocos Plate, RP, Riviera Plate, and CaP, Caribbean Plate. The triangles represent the volcanoes with eruptions in historic times in Mexico (De la Cruz-Reyna, 2008).

It was proposed (Zobin et al., 2008; 2009) that the low-frequency phase was generated by the movement of fragmented magma in the conduit and the duration of this low frequency phase could indicate the depth of magma fragmentation.

This chapter discusses the seismic signals associated with the Popocatepetl Vulcanian explosions together with the acoustic signals generated during these explosions. In recent years the importance of monitoring volcanoes with acoustic sensors has been demonstrated (Johnson et al., 2003a; 2003b).

Acoustic studies help to better understand the volcanic processes occurring during magma degassing and explosions. The use of infrasound has helped in the understanding of eruption physics, the location of various types of volcano infrasound sources, the inference of vent and conduit geometries, and/or quantifying the outflux of volcanic materials (Johnson et al., 2003a; 2003b; 2004; Johnson and Ripepe, 2011). The Green's function of the atmosphere, through which the infrasound waves pass from the source to the microphone, is much simpler than that for the propagation of seismic waves in the ground (Johnson et al., 2003a).

Therefore, infrasound records are good indicators for understanding eruption mechanisms, their characteristics directly reflect those of the source, when the station is close to the source (Yokoo et al., 2009).

Also, these signals exhibit low atmospheric scattering/ dissipation and experience predictable echoing, site, or weather dependent effects. Infrasound is also largely unaffected by cloud cover and does not rely on line of sight view of eruptive vents. Some variations in infrasound traces recorded at a local network can be attributed to explosive source directivity, near-vent crater morphology (e.g., wall echoes), or site effects resulting from topography near of the station (Jones and Johnson, 2011).

It has been proposed that Vulcanian explosions form through the rupture of an impermeable cap within the vent at a shallow level when the gas pressure increases in the pore spaces beneath the cap to a level that exceeds the lithostatic pressure above. There are many examples of seismic-acoustic explosion signals recorded at andesitic volcanoes, such as Sakurajima (Garces et al., 1999; Morrissey et al., 2008; Yokoo et al., 2009), Tungurahua (Johnson et al., 2003b; Ruiz et al., 2006), Pichincha (Johnson et al., 2003b), Soufrière Hills (Ripepe et al., 2010), Fuego (Johnson et al., 2004), just to mention a few. The analysis of seismic and acoustic information presented in these publications resulted in a better understanding of the explosive phenomena.

The aim of our study is to improve our knowledge of the generation mechanism of Vulcanian explosions at Popocatepetl through seismic and acoustic information about the largest explosions that occurred during the period from 2008 to 2012 (Figure 2). A summary of the 13 explosions studied is shown in Table I. The Vulcanian explosions were short and discrete, with duration of a few minutes. Some explosions were large enough (especially the explosion of 20 November 2011) to be heard over large distances, as far as the city of Cuernavaca, 64 km from the crater.

1.1. Data

The recent seismic network installed around Popocatepetl volcano consists of seven short period stations, equipped with triaxial seismometers Mark L4C, T=1 sec (PPP, PPX, PPJ, PPQ, PPM, PPT and PPC).

Additionally, the network has been augmented with four triaxial broadband seismometers (three Guralp CMG-40TD of 0.02-30s instruments PBM, PBX, PBC), and one CMG-40T of 0.02-60s (PBP).

The acoustic sensors were co-located at two stations PSM and PSC and consisted of Panasonic microphone WM-034BY, showing an approximate flat response between ~ 1 to ~ 5 Hz and a sensitivity of 56 mV/Pa. Two acoustic-seismic stations, utilized for this study, were Tlamacas (PPM, PBM and PSM) and Colibrí (PPC, PBC and PSC), located at 4.8 km and 8.4 km from the crater, respectively.

Some explosions were observed only at one station, due to technical problems with the other station. Four explosions (No. 3, 4, 9 and 10 in Table 1) were clipped at station PSM at the beginning of the signal due to the low resolution of the analogical-digital converter or due to the strength of the explosion and the response of the sensor.

In Figure 3, the location of the current network is shown. The seismic and acoustic data were sampled at a frequency of 100 Hz. All these signals arrived in real time to the processing centres located in Mexico city at the National Disaster Prevention Centre (CENAPRED) and at the Geophysical Institute of the National Autonomous University of Mexico (UNAM), to be analysed and processed.

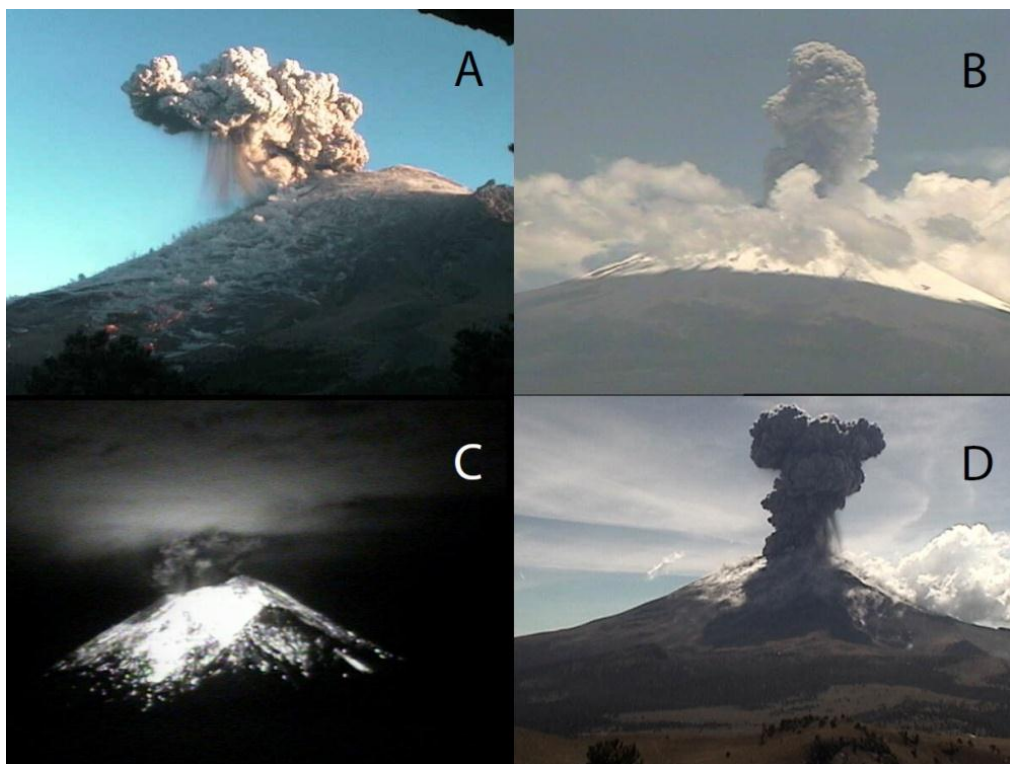


Figure 2. Vulcanian explosions at Popocatepetl volcano. A) 9 March 2008, B) 7 June 2010, C) 26 September 2011, and D) 20 November 2011. Photos were taken from <http://www.cenapred.unam.mx/es/Instrumentacion/InstVolcanica/MVolcan/>.

Table 1. List of 13 explosions at Popocatépetl volcano

Nos	Date (seismic time, UTC)	Volcanic Activity
1	08-feb-2008, 03:37:51	Explosion with an ash column.
2	09-mar-2008, 00:32:22	Explosion with a column of ash of 2 km and incandescent fragments to the northeast of the volcano edifice.
3	23-mar-2009, 08:12:04	Explosion with a column of 1.8 km to the southeast and incandescent fragments propelled a short distance from the crater.
4	26-may-2010, 02:24:45	Explosion with an ash column.
5	07-jun-2010, 18:07:41	Explosion with a column of ash, which reached 3 km altitude to the northeast
6	10-jun-2010, 03:43:47	Explosion with ash column of 1.5 km of altitude to the west.
7	11-jun-2010, 07:33:04	Explosion with ash column of 1 km of altitude.
8	03-jun-2011, 11:41:00	Exhalation with explosive component with an ash column of 3 km, to different directions.
9	26-sep-2011, 07:59:00	Large explosion with a column of ash of 1.5 km to the north and incandescent ballistics covered the entire upper flanks of the volcano.
10	20-nov-2011, 18:00:59	Large explosion with a column of ash of 4.5 km to the east and incandescent ballistics expelled from the crater. Explosion heard in Cuernavaca located 64 km from the crater.
11	25-jan-2012, 16:10:04	Explosion with ash, reaching a height of 6 km to the northeast.
12	14-apr-2012, 03:20:32	Explosion with few ballistics expelled from the crater.
13	14-apr-2012, 03:35:31	Explosion with few ballistics expelled from the crater to the east.

Note: The description of volcanic activity was taken from <http://www.cenapred.unam.mx/cgi-bin/popo/reportes/consulta.cgi>.

II. METHODOLOGY

II.1. Spectral Studies

Spectrograms of the seismic and acoustic signals were created from Fourier amplitude spectra calculated within fixed time windows of 5.12 sec, overlapping by up to 50%. The Fourier amplitude spectrums were calculated for the entire signal.

II.2. Reduced Pressure

The acoustic records allow us to estimate the reduced pressure of the explosion. For acoustic sensors deployed within about 10 km from a volcano, the pressure variations are often assumed to be linear acoustic waves (Johnson and Ripepe, 2011).

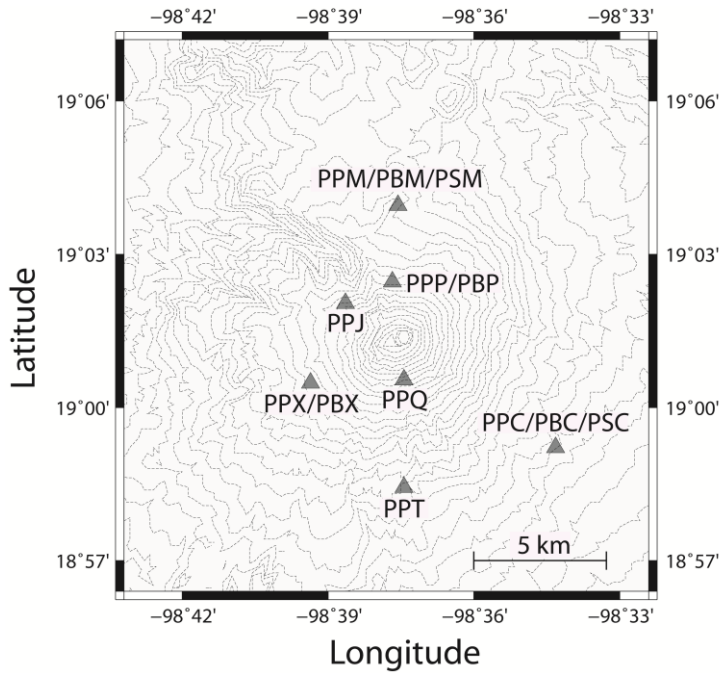


Figure 3. Seismic and acoustic monitoring network. All the seismic stations have short period seismometers (PPM, PPP, PPJ, PPX, PPQ, PPT and PPC). Additionally, the stations PBM, PBP, PBX and PBC have broadband sensors also. Two acoustic sensors were installed, one at PSM and the other PSC seismic stations.

Therefore, the maximum amplitudes of acoustic signals may be taken as a value of pressure P at the station. For the comparison of the pressure amplitudes from different volcanoes and different eruptions, the reduced pressure was calculated (equation 1).

This equation allowed the transformation of the pressure P recorded at distance r to a reduced pressure P_{red} , that would have been recorded at a distance r_{red} from the volcano (Johnson and Ripepe, 2011).

$$P_{red} = P \frac{r}{r_{red}} \quad (1)$$

The reduction distance r_{red} is arbitrarily equal to 1 km (Johnson et al., 2004). Although variations in the amplitude of pressure can depend on the local meteorological conditions, this factor was not taken into account in this study because of the proximity of the sensors (< 9 km).

II.3. Seismic and Acoustic Energies

To calculate the seismic energy for the entire signal, equations (2) and (3) proposed by Boatwright (1980) for an isotropic source located at the top of a homogeneous half space were used.

$$E_{Seismic} = 2\pi r^2 \rho_{earth} c_{earth} \frac{1}{A} \int S^2 U(t)^2 dt \quad (2)$$

$$A(r) = e^{(-\pi f r)/(cQ)} \quad (3)$$

here r is the approximate distance between the source and the seismic station, c_{earth} is the P wave velocity, S is the site effect, $U(t)$ is the seismic amplitude, ρ_{earth} is the density of the medium, A is the attenuation, f is the predominant wave frequency and Q is the quality factor. The values used were for the seismic velocity of $c_{earth}=3500$ m/s (Valdés-González et al., 1995), site effect $S=1$, density $\rho_{earth}=2000$ Kg/m³, frequency $f=2$ Hz and quality value $Q=10$ (Johnson and Aster 2005). These equations were used by Johnson and Aster (2005) and Johnson et al. (2005) to calculate the seismic energy of explosions at Karymsky (Kamchatka), Erebus (Antarctica) and Santiaguito (Guatemala) volcanoes.

In the case of the acoustic energy, the equation (4) proposed by Johnson and Aster (2005), was used:

$$E_{Acoustic} = \frac{2\pi r^2}{\rho_{atmos} c_{atmos}} \int \Delta P(t)^2 dt \quad (4)$$

where r again is the approximate distance between the source and the acoustic station, ρ_{atmos} is the density of the atmosphere, c_{atmos} is the acoustic velocity in the atmosphere and $\Delta P(t)$ is the excess pressure observed with the acoustic sensor. For the acoustic energy, the atmospheric density $\rho_{atmos}=1.2$ kg/m³, and the acoustic velocity in the atmosphere $c_{atmos}=340$ m/s (Ripepe et al., 2001; Johnson and Aster, 2005).

II.4. Volcanic Acoustic–Seismic Ratio VASr

The Volcanic Acoustic–Seismic Ratio VASR was calculated using the equation (5) proposed by Johnson and Aster (2005):

$$\eta = \frac{E_A}{E_S} \quad (5)$$

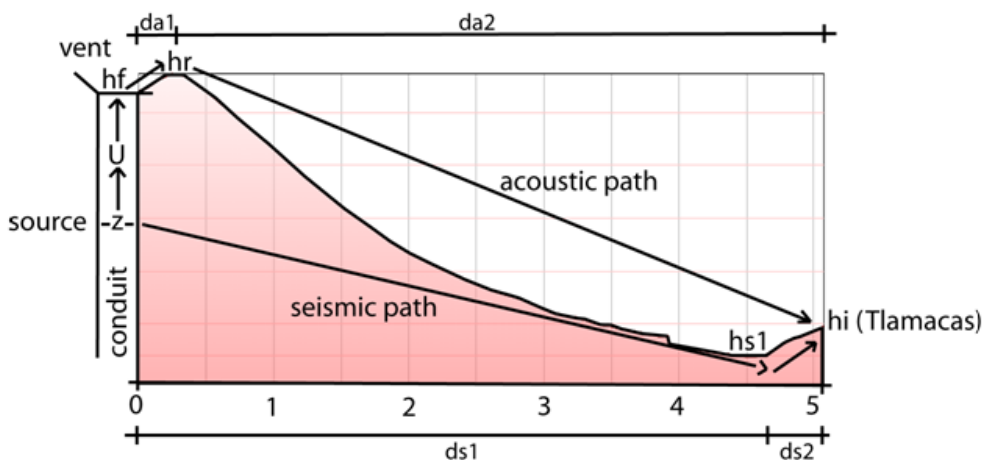
where E_A is the acoustic energy and E_S is the seismic energy. The VASR offers the possibility of improving our understanding of the seismo-acoustic source, such as physical characteristics of magma or the conduit geometry.

The use of VASR values allows an inter-comparison of explosive behaviour at different volcanoes, as well as the examination of changing conditions within a suite of explosive events at a single volcano (Johnson and Aster, 2005).

II.5. Estimation of the Depth of Acoustic-Seismic Source

The approximate depth of the acoustic-seismic source was estimated based on the methodology proposed by Ruiz et al. (2006). This methodology was slightly modified for our study. It was assumed that the seismic and acoustic source was generated at the same time and place. The source depth below the crater was taken as z .

From this depth, material in the gas and solid phase ascended at velocity (U) through the conduit, reaching the crater floor (altitude h_f) at time $T_0 + z/U$. When the acoustic signal reaches the crater rim (altitude h_r), it travels to the acoustic sensor situated at an altitude h_i (Figure 4). In the case of the seismic signal, the wave travels from z to a point at altitude (h_{s1}) and then up to the seismic station situated at an altitude h_i .



Modified from Ruiz et al., 2006.

Figure 4. Schematic paths of acoustic and seismic signals from volcano to Tlamacas station, where h_i is the altitude of Tlamacas station, h_f is the altitude of the crater floor, h_r is the altitude of crater rim, d_{hi} the horizontal distance from the conduit to station and $d_{hi} = d_{s1} + d_{s2} = d_{a1} + d_{a2}$, z is the depth of the acoustic-seismic source, and U the velocity of the gas and solid material ascent inside the conduit.

The following equations (6, 7, and 8), taken from Ruiz et al. (2006), were used to estimate the travel times of seismic and infrasonic waves to the Tlamacas station, situated on a hill:

$$t_{pi} = \frac{(\sqrt{d_{s1}^2 + (h_f - h_{s1} - z)^2} + \sqrt{d_{s2}^2 + (h_i - h_{s1})^2})}{C_{earth}} \quad (6)$$

$$t_{aci} = \frac{\sqrt{d_{a1}^2 + (h_r - h_f)^2} + \sqrt{d_{a2}^2 + (h_r - h_i)^2}}{C_{atmos}} \quad (7)$$

The distances d_{s1}^2 , d_{s2}^2 , d_{a1}^2 and d_{a2}^2 are shown in Figure 4. The time difference between arrivals is the following:

$$\Delta T_i = \frac{z}{U} + t_{aci} - t_{pi} \quad (8)$$

here t_{pi} is the travel time of the first seismic pulse from the source to the station i , c_{earth} is the seismic (P wave) velocity in the volcano edifice, t_{aci} is the travel time of acoustic wave from the crater to the station i and c_{atmos} is the velocity of infrasound in the atmosphere.

In the case of the Colibri station, situated on the plain, the equation (6) was simplified as (9):

$$t_{pi} = \frac{\sqrt{d_i^2 + (h_f - h_i - z)^2}}{c_{earth}} \quad (9)$$

here d_i is the horizontal distance between the source and the seismic station.

III. RESULTS

Seismic signals associated with Vulcanian explosions can be interpreted as a result of liberation of elastic energy due to the rupture of a seal or cap due to high pressures beneath. The waveform of these events at Popocatepetl is complex, due to many contemporaneous processes.

As was mentioned by Sanderson et al., (2010) for the Santiaguito dome complex, Santa Maria volcano, Guatemala, these processes can be associated with volumetric changes in a specific region of the conduit due to mass transport to the surface, response of the pressurized gas flow and/or sub-vertical motion of the dome's upper surface, response to gas/pyroclastic emissions, or fracturing of rock and/or ballistic impact from fallout and associated pyroclastic flows.

The raw seismic signals at Popocatepetl are composed of the initial low frequency phase followed by the high-amplitude and high-frequency signal (Figure 5A). Figure 5B shows the spectrogram and Fourier spectra for the 26 September 2011 explosion (Figure 5C). The spectrogram illustrates a wide range of frequencies presented in the signal. The source mechanism is complex, but certain contributions can be defined. For example, the high frequencies are associated with incandescent ballistics that were thrown and landed on the volcano flanks.

In the case of the acoustic emissions, as usual, the signals are simpler, generally presenting the characteristic N shape, and sometimes finishing with tremor, lasting for a few minutes. This is thought to be a consequence of gas emission (Figure 6A). In Figure 6B the spectrogram and the amplitude spectrum of an acoustic signal are shown (Figure 6C).

The peak frequencies of both seismic and acoustic signals were calculated for the entire signals. For the seismic data, they varied between 0.89 and 1.41 Hz at PBMV station, and between 0.43 and 1.81 Hz at PBCV station (Table II). For the acoustic signals, they varied between 0.23 and 1.12 Hz at PSM station, and between 0.24 and 1.63 Hz at PSC station. Some acoustic signals were saturated at PSM.

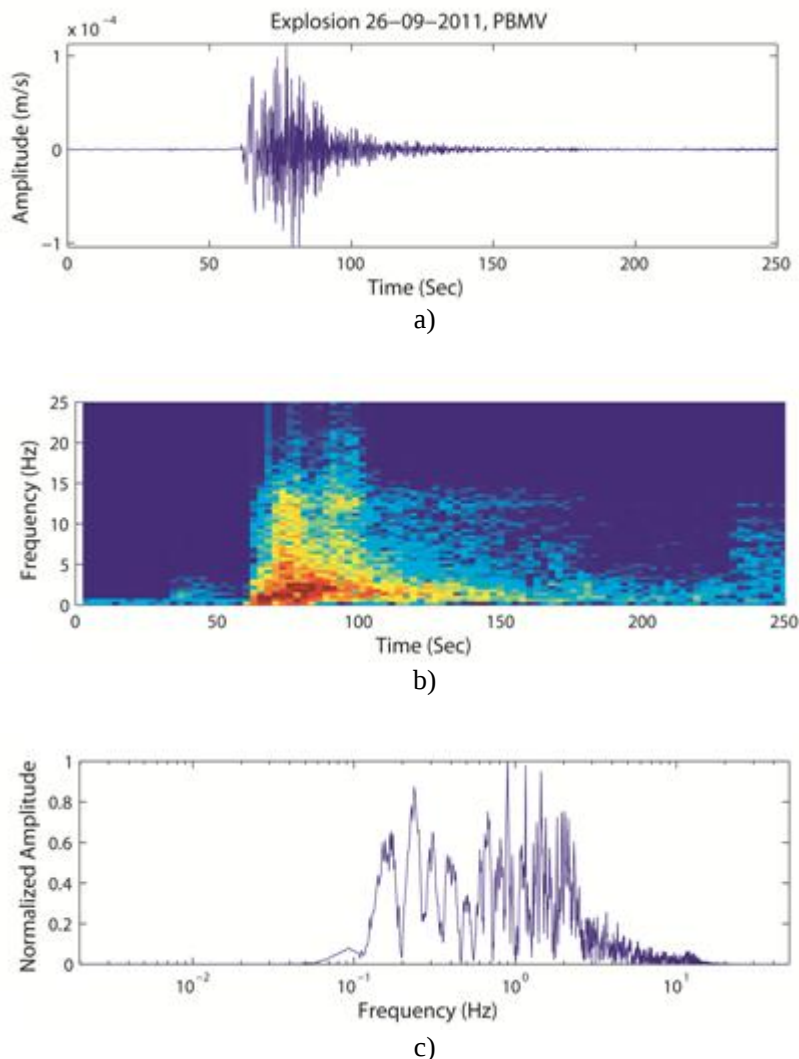


Figure 5. A) Seismic signal of the 26 September 2011 explosion recorded by broad-band station (PBMV). The start of the trace is at 07:58:00 (GMT). B) Spectrogram calculated with fixed time windows of 5.12 sec and overlapping by up to 50%. C) Normalized amplitude spectrum of the seismic signal. The spectrum was calculated from the entire signal.

Four spectral bands between 30 sec and 50 Hz were calculated for the vertical component of the seismic signals. The Guralp CMG-40TD sensors in PBMV and PBCV have a flat response between 30 sec and 100 Hz. In spite of the sampling frequency was 100 Hz, we can only observe up to 50 Hz, which represents the Nyquist frequency associated with aliasing. The data were deconvolved and filtered with a two-pole minimum phase causal Butterworth filter. The bands utilized were: Very-Long-Period (VLP, 30-10 sec), Long-Period (LP, 10-1 sec), Short-Period (SP, 1-10 Hz) and High-Frequency (HF, 10-50 Hz). VLP signals were obtained only for three explosions at Tlamacas station (No. 9, 10 and 11) and for explosion No. 10 at the Colibrí station. The VLPs were observed at all three components of the seismic signals. In Figure 7, four bands of the seismic signal and the acoustic signal are shown for the explosion occurring on 20 November 2011 (No 10).

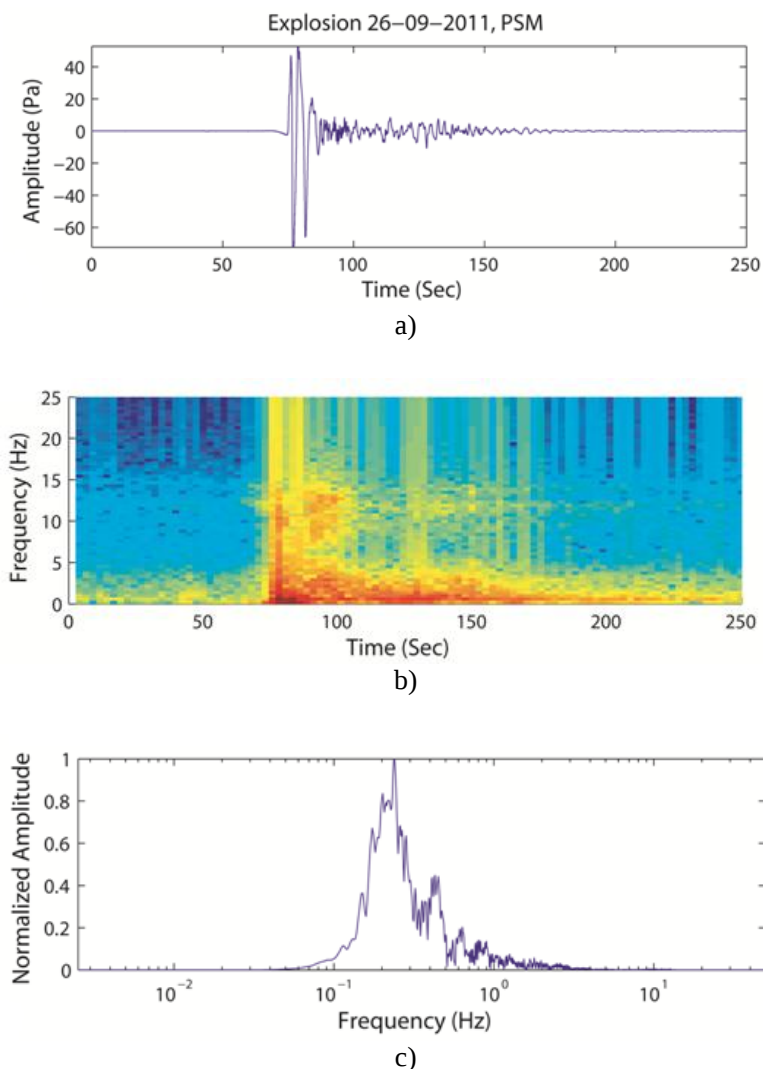


Figure 6. A) Acoustic signal of the 26 September 2011 explosion recorded at station PSM, the start of the trace is at 07:58:00 (GMT). B) Spectrogram calculated with fixed time windows, overlapping by up to 50%. C) Normalized amplitude spectrum of the acoustic signal. Again the spectrum was calculated from the entire signal.

Figure 7A shows that for the VLP component we have long-period vibrations without any special phases. This VLP is similar to that described by Arciniega et al., (1999). In the case of the LP band (Figure 7B), the first pulse of low-frequency energy is clearly seen. It is followed by higher-frequency large-amplitude phase.

These low frequency pulses were observed also by Zobin et al. (2008) for the seismic records of many explosions of Popocatepetl occurring during 1998-2000, and his methodology resulted in a calculation of the counter force of the explosions from these phases.

For the SP band (Figure 7C), the amplitude of the signal is slightly less than for the LP band. The HF band is characterized by the largest amplitude of the seismic signals generated by this explosion recorded in 74 seconds after the first arrival (Figure 7D).

Table 2. Spectral characteristics of the events recorded at (PPMV/PBMV), PSM, (PPCV/ PBCV) and PSC. Some acoustic signals were clipped and marked with (c)

No	Date and time, dd-mm-yyyy, hh:mm:ss	Seismic peak frequency recorded at PPMV/PBMV (Hz)	Acoustic peak frequency recorded at PSM (Hz)	Seismic peak frequency recorded at PPCV/PBCV (Hz)	Acoustic peak frequency recorded at PSC (Hz)
1	08-02-2008, 03:37:51	0.95	1.12	-	-
2	09-03-2008, 00:32:22	1.16c	0.28	1.71	-
3	23-03-2009, 08:12:04	1.03c	0.23	1.14	0.43
4	26-05-2010, 02:24:45	-	-	1.81	0.47
5	07-06-2010, 18:07:41	-	-	0.94	0.25
6	10-06-2010, 03:43:47	1.33	-	1.44	1.03
7	11-06-2010, 07:33:06	1.41	-	1.40	0.62
8	03-06-2011, 11:41:00	-	-	1.79	1.63
9	26-09-2011, 07:59:00	0.89c	0.23	1.81	0.24
10	20-11-2011, 18:00:59	0.89c	0.23	0.43	0.43
11	25-01-2012, 16:10:04	0.89	0.24	0.96	0.25
12	14-04-2012, 03:20:32	-	-	1.98	0.49
13	14-04-2012, 03:35:31	-	-	1.73	0.54

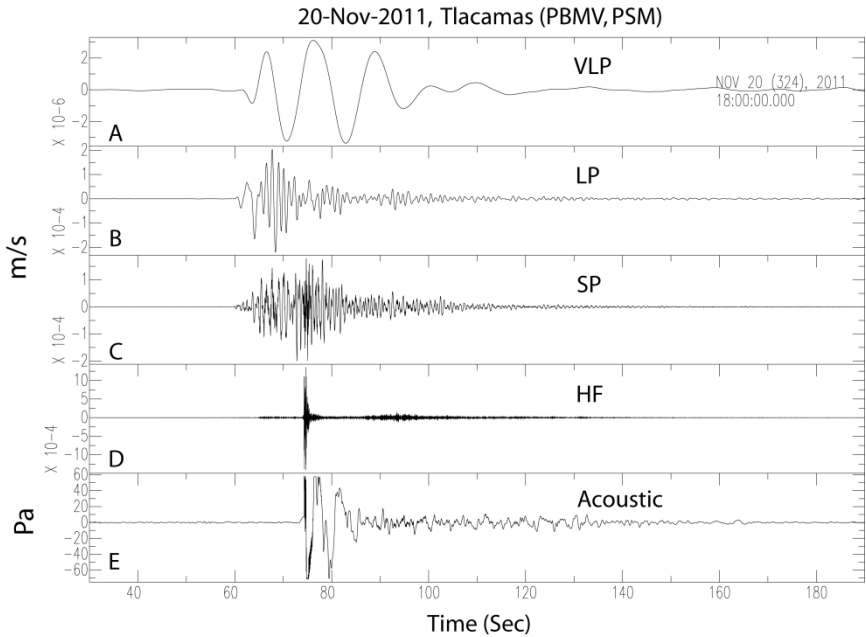


Figure 7. Seismic and acoustic signals of the 20 November 2011 explosion at PBMV and PSM. A) Very-Long-Period seismic band (VLP, 30-10 sec), B) Long-Period seismic band (LP, 10-1 sec), C) Short-Period seismic band (SP, 1- 10 Hz), D) High-Frequency seismic band (HF, 10-50 Hz), E) Acoustic signal from the station of PSM, this signal was clipped in its start. Note that the HF and acoustic signals register the arrival of the acoustic-airwave of the explosion at the same time.

This large-amplitude HF pulse is associated with the air wave of the explosion evidenced by the contemporaneous arrival of the acoustic signal (Figure 7E). This explosion was heard up to 64 km from the crater. The air-phase in the HF seismic band is followed by a sequence of vibrations supposedly associated with the fall of ballistic material.

Estimation of the cumulative energies of the seismic and acoustic signals is illustrated for the 26 September 2011 explosion (Figure 8). Both signals, acoustic and seismic, were filtered with a pass-band two-pole Butterworth filter between 0.1 and 12 Hz. Figure 8 (B and D) show the seismic and acoustic cumulative energy derived from the signals.

The seismic energy of the explosion was calculated as the maximum of accumulated energy. The energy of 9 explosions was estimated using the records of Colibri station, and the energy of 6 explosions was estimated using the records of Tlamacas station. Only 4 events were analyzed using the records of both stations. All energy estimations are presented in Table III together with the values of reduced pressure and VASR.

The values of reduced pressure varied from 7.8 to 393.4 Pa at the PSM station and from 2.1 to 373.9 Pa at PSC station. The seismic energy varied between 4.5 and 633.4 MJ for PPMV/PBMV, and between 0.9 and 1691.9 MJ for PPCV/PBCV. For the acoustic signals, the energy varied between 1.1 and 6803.2 for PSM and between 0.25 and 3448.3 MJ for PSC. The VASR values for Tlamacas station were between 0.19 and 96.5, and for Colibrí station between 0.05 and 16.8.

The largest explosion occurred on 20 November 2011. Its seismic energy was 633.4 MJ at PBMV station and of 1691.9 MJ at PBCV station; its acoustic energy was 6803.2 MJ at PSM and 3448.3 MJ at PSC station. The reduced pressure for this event was of 380.2 at PSM and of 373.9 at PSC.

The mean seismic energies for the 4 events recorded at two stations ranged from 182 to 1162.2 MJ and from 330.6 to 5125.8 MJ for the acoustic energies.

In the case of the VASR, values are between 0.73 and 16.8. The approximate precision of our estimation of the quantitative parameters may be calculated from the comparison of the values obtained at two stations for the same events.

For the reduced pressure the precision is equal to ± 72.25 Pa; for seismic energy, it is of ± 1.09 log units (MJ); for acoustic energy it is of ± 0.47 log units (MJ), and for VASR values, of ± 39.92 .

The delay times of the acoustic signal with respect to the seismic signal are shown in Table IV. For Tlamacas station the times are between 12.74 and 15.08 sec. For the delay times, observed at the Colibrí station, the values are between 20.40 and 24.05. For this station, the three highest values are 38.57, 71.91 and 123 sec.

Theoretical curves for time delay between the seismic and acoustic phases were calculated from equation (8) considering a variation in the seismic velocity c_{earth} in the region of Popocatepetl of between 2 and 3.5 km/s (Valdés et al., 1995). The acoustic velocity c_{atmos} was fixed at 340 m/s (Ripepe et al., 2001; Johnson and Aster, 2005).

The velocity of ascent of the magmatic products in the conduit U was taken between 50 and 100 m/s (Ripepe et al., 2001; Ruiz et al., 2006). With these data, different curves were calculated for both stations (Figure 9).

The depth of the source was calculated from the comparison of the theoretical and observed data. The maximum depth was estimated between 200 and 220 meters (Figure 9) below the crater that corresponds to the seismic velocity c_{earth} of 2 km/s and a velocity of pyroclastic material ascent U of 100 m/s.

Table 3. Energy characteristics of the seismic and acoustic signals of explosions recorded at the stations of Tlamacas (PPMV, PBMV, PSM) and Colibrí (PPCV, PBCV, PSC)

No	Date	Pressure reduced P_{red} (Pa) PSM	Pressure reduced P_{red} (Pa) PSC	Seismic energy E_s (MJ) PPMV/PBMV	Acoustic energy E_a (MJ) PSM	Seismic energy E_s (MJ) PPCV/PBCV	Acoustic energy E_a (MJ) PSC	VASR Tlamacas	VASR Colibrí
1	08-feb-2008	9.0	x	5.7	1.1	x	x	0.19	x
2	09-mar-2008	118.6c	x	36.0	562.6c	x	x	15.6	x
3	23-mar-2009	103.0c	138.4	4.5	442.3c	892.7	218.9	96.5	0.24
4	26-may-2010	7.8	6.3	x	x	7.8	0.4	x	0.05
5	7-jun-2010	x	14.0	x	x	2.5	14.4	x	5.2
6	10-jun-2010	x	2.1	50.0	x	1.5	0.25	x	0.16
7	11-jun-2010	x	4.0	32.2	x	0.9	0.58	x	0.64
8	03-jun-2011	x	31.65	x	x	6.9	116.8	x	16.8
9	26-sep-2011	393.4c	351.7	115.3	6609.4c	248.7	2141.4	57.2	8.6
10	20-nov-2011	380.2c	373.9	633.4	6803.2c	1691.9	3448.3	10.7	2.0
11	25-jan-2012	377.0	171.4	461.0	2962.9	1619.3	423.9	6.4	0.26
12	14-apr-2012	x	37.7	x	x	41.8	150.0	x	3.5
13	14-apr-2012	x	195.3	x	x	111.2	759.3	x	6.8

Table 4. Delay times between the seismic and acoustic phases recorded at the stations of Tlamacas and Colibrí

No.	Date	Delay Times (Tlamacas)	Delay Times (Colibrí)
1	08-feb-2008	14.33	x
2	09-mar-2008	14.44	x
3	23-mar-2009	15.08	23.94
4	25-may-2010	x	23.59
5	7-jun-2010	x	29.65
6	10-jun-2010	x	38.57
7	11-jun-2010	x	71.91
8	3-jun-2011	x	123
9	26-sep-2011	13.77	22.60
10	20-nov-2011	13.81	24.05
11	25-jan-2012	12.74	20.40
12	14-apr-2012	x	x
13	14-apr-2012	x	21.0

IV. DISCUSSION

Vulcanian explosions are complex events that involve many different processes occurring within the volcanic conduit and associated with the outlet of volcanic products. This complexity of eruption processes is reflected in different frequency bands of the seismic signals.

The VLPs are supposed to be associated with volumetric changes in a specific region of the conduit, due to mass transport to the surface (Giudicepietro et al., 2009) or with action of

a recoil force from a volcanic jet during an explosion (Chouet, 2003). Other studies have proposed that the depth of the source of the VLPs at Popocatépetl is about of 1-1.5 km below the crater, (Arciniega et al., 1999; Chouet et al., 2005). The study of the VLP band of the Vulcanian explosions of Popocatépetl presented in this work shows also that only strong explosions beginning from the seismic energy of 115 MJ at PBMV and of 1691 MJ at PBCV could generate this kind of event.

In the case of the LP band, the occurrence of a low frequency phase in this band with duration between 3 and 15 seconds was observed by Zobin et al. (2006; 2008) for Vulcanian explosions of Volcán de Colima and Popocatépetl. The nature of this phase has been debated: Zobin (2009) considered that it is generated by the ascent of fragmented material inside the conduit.

Although, in the case of Vulcanian explosions at Volcán de Colima, Varley et al. (2010) mentioned that after of the initial rupture of the impermeable plug there is a rapid pressure drop and degassing of the magma beneath. This could be related to the initial low-frequency part of the seismic signal.

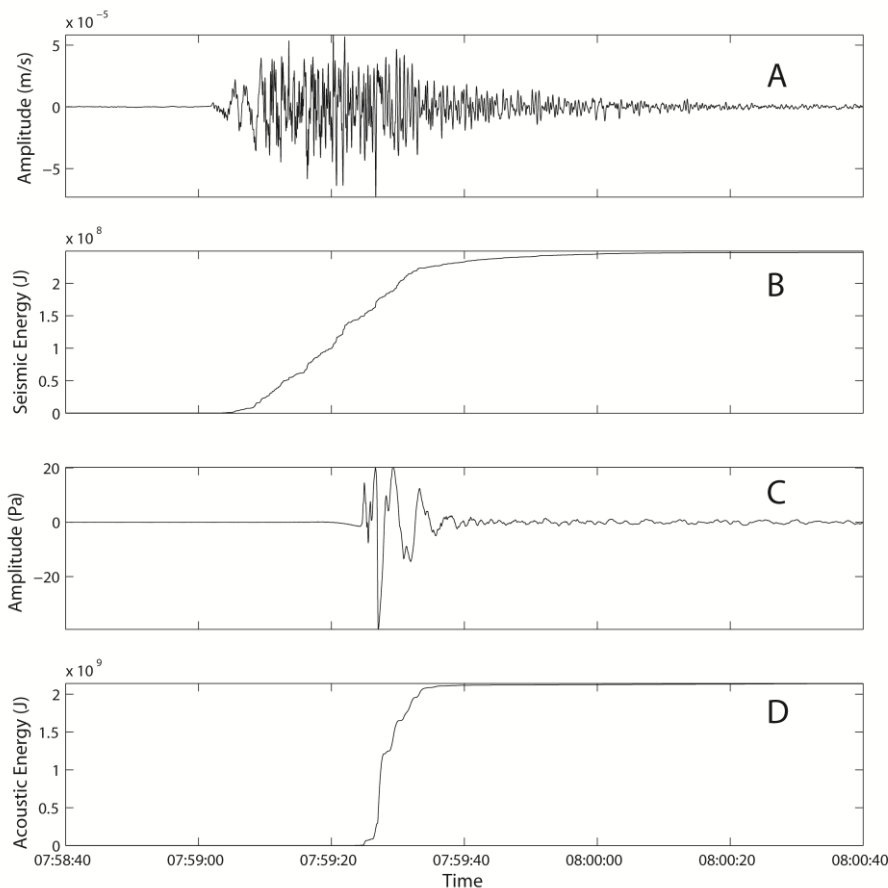


Figure 8. A) Seismic signal at PBCV of the explosion on 26 September 2011, B) Cumulative seismic energy, C) Acoustic signal at PSC, D) Cumulative acoustic energy. All the traces were filtered between 0.1 and 12 Hz.

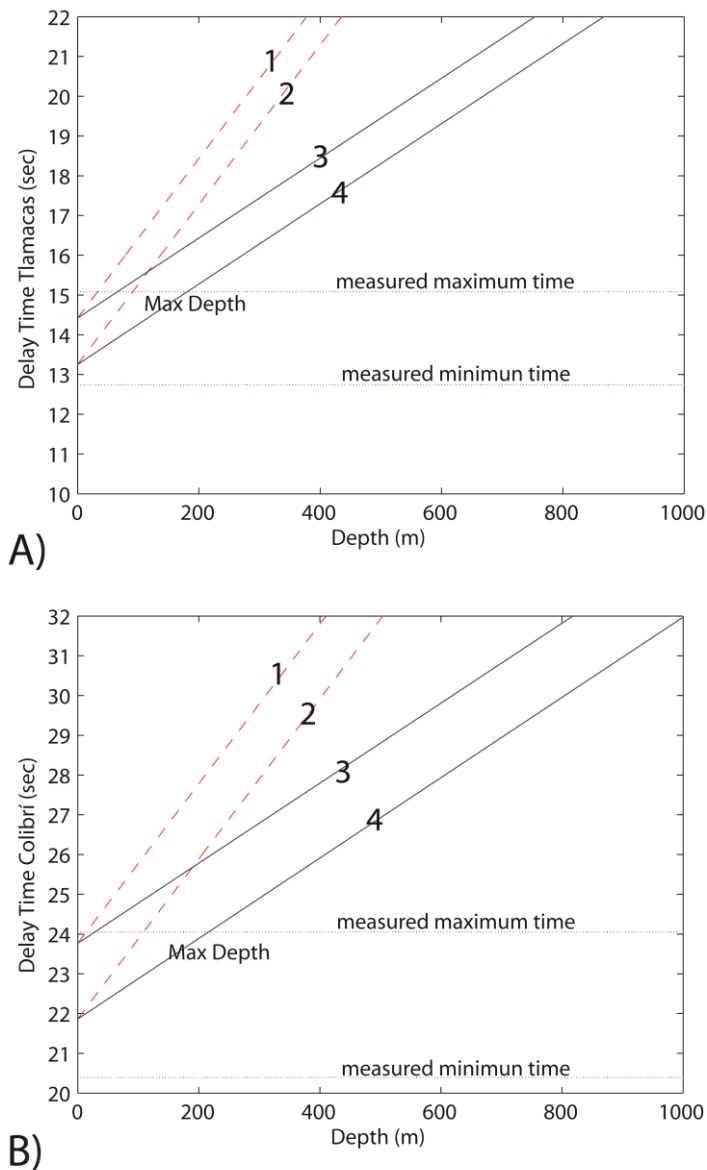


Figure 9. A) Theoretical delay times for Tlamacas vs depth (z), using $U = 50$ m/s and $c_{\text{earth}} = 3.5$ km/s (line 1), $U = 50$ m/s and $c_{\text{earth}} = 2.0$ km/s (line 2), $U = 100$ m/s and $c_{\text{earth}} = 3.5$ km/s (line 3) and $U = 100$ m/s and $c_{\text{earth}} = 2.0$ km/s (line 4). B) The same for the Colibrí acoustic station.

Considering the SP band, is supposed that this spectral band is generated due to the flow of magmatic material and gas in the conduit before their exit of the crater. This effect was observed for such volcanic events as the gas exhalations or volcanic tremor (Arciniega et al., 2000; Arciniega et al., 2003). It is likely that the HF band includes several components: the fall of material on the surface of the volcanic edifice, the occurrence of pyroclastic flows and also the airwave of the explosion, as could be observed in Figure 7. The predominant

frequencies, observed in the seismic signals, are probably related to the size of the source, the composition of the magma and the characteristics of the conduit.

In the case of the acoustic signals, these events are generated by the liberation of energy into the atmosphere. The microphones, installed at Popocatepetl, had a narrow band of spectral response within 1-5 Hz, so it was not possible to analyse different bands through filtering, similar to Johnson et al. (2009). The relatively low-frequency of the acoustic signals, approximately between 0.2 and 1.6 Hz, was probably related to the dimensions of the source. It has been shown that the source dimensions are related to the infrasonic wavelengths (17m for 20Hz to 340m for 1Hz; Johnson et al. 2003). The recorded frequencies were similar to those observed for other Vulcanian explosions at volcanoes Augustine (Petersen et al., 2006), Tungurahua (Ruiz et al., 2006; Fee et al., 2010), and Sakurajima (Morrissey et al., 2008).

The data presented in Table III allow us to describe the variety in the source parameters of Vulcanian explosions. The values of reduced pressure, obtained at two stations, were very similar, with a difference not greater than 41.7 Pa; only for the explosion on 25 January 2012 the difference reached of 205 Pa, probably due to an anisotropic source. Some explosions were clipped at the PSM station; therefore the pressure of these explosions was underestimated. The maximum value of the reduced pressure between 373 Pa at Colibrí station was estimated for the explosion occurring on 20 November 2011. This value can be compared with 200 Pa recorded for explosions at Erebus and Tolbachick volcanoes (Firstov and Kravchenko, 1996), and 150 Pa recorded at Tungurahua (Johnson and Ripepe, 2011).

The values of the seismic and acoustic energies varied for two stations. In the case of Tlamacas, installed on the hill, these values ranged from 4.5 to 633.4 MJ for the seismic energy and from 1.1 to 6803.2 MJ for the acoustic energy. The maximum values for acoustic energy were underestimated due to the saturation of the signals. For the Colibrí station, installed on the plain, the values ranged from 0.9 to 1691.9 MJ for seismic energy, and 0.25 to 3448.3 MJ for acoustic energy. The difference in the acoustic and seismic energies obtained at the two stations is probably due to different attenuation of the signals on the way from the volcano to the stations.

Our estimations of the seismic energy for Vulcanian explosions are higher with respect to those reported for the Erebus and Karymsky explosions (0.0011- 11 MJ, Johnson et al., 2005) and calculated with the same methodology. The values for the acoustic energies for Vulcanian explosions of Popocatepetl are comparable with the estimations of this parameter for the same type of explosions at Sakurajima (400 MJ; Garcés et al., 1999), Augustine (200 MJ-4000 MJ; Petersen et al., 2006) and Tungurahua (190 MJ; Johnson and Ripepe, 2011).

The VASR values estimated for the seismic and acoustic records at both stations were of the order of 0.05 to 96.5. This variety in VASR allowed us to discriminate the explosions as small (VASR < 1), moderate (VASR between 1 and 10), and large (VASR >10). The VASR demonstrated a tendency to increase with an increase of the seismic and acoustic energy of the explosions. In the case of some explosions recorded with the saturated signals at Tlamacas (Nos. 2, 3, 9 and 10 in Table III) values of VASR are between 10.7 and 96.5 and may be considered as subestimated in this station. Five explosions (Nos 2, 3, 9, 10, 9 and 13 from Table I) were accompanied with the fall of the incandescent fragments from the crater. They were characterized by the values of VASR from 15.6 to 96.5 at Tlamacas and from 0.24 to 8.6 at Colibrí.

In Figure 10, two kinds of explosions can be observed: five small explosions (acoustic energy < 20 MJ) accompanied with gas-and-ash plumes and eight moderate explosions

(acoustic energy > 200 MJ) associated with exit of incandescent ballistics. The small explosions have mainly low VASR values (0.1-1), while the moderate explosions have moderate to high values of VASR between 1 and 100. That means that the seismic energy of moderate explosions is smaller than the acoustic energy of these explosions. Johnson and Aster (2005) considered that the size of VASR may be related to the geometry of the conduit or the size of the source where the explosions were originated. They proposed that a magmatic flow within a wide, smooth-walled conduit is expected to experience less viscous resistance than flow through a narrow tephra-choked conduit. When the conduit is a short and wide, the resistance is high, and the seismic efficiency reduced, increasing the VASR. In the case of the size of the source, a large acoustic source region may result in diminished radiated acoustic efficiency; therefore, small sources can generate high acoustic energy and high VASR values. According to the model of Johnson and Aster (2005), the moderate and high values of VASR obtained for our moderate-size explosions could be associated with a short and wide conduit and/or a small source region. The mean delay times between seismic and acoustic signal onsets, obtained in our study (Table IV), indicate that the depths of the acoustic-seismic sources are mainly shallow, varying between 0 and 220 m beneath the crater floor, occurring beneath or inside the lava domes.

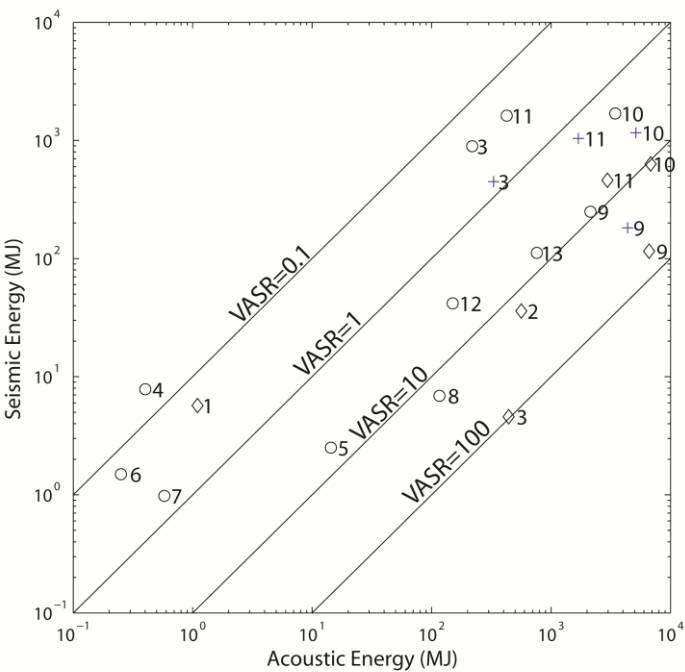


Figure 10. Relation between the seismic and acoustic energies of Vulcanian explosions at Popocatepetl. The lines of equal VASR are shown for 0.1, 1, 10, and 100. Diamonds represent data obtained at the Tlamacas station, circles, at Colibrí station, and crosses are the mean values. Two groups of explosions could be separated according to the difference in the acoustic energy: small than 20 MJ) and moderate explosions accompanied by ballistic fragments thrown from the explosions (left side, less crater (right side, over 100 MJ).

Similar shallow depths for Vulcanian explosions have been observed also at other volcanoes, e.g. Tungurahua (Ruiz et al., 2006), Stromboli (Chouet et al., 1997; Ripepe et al., 2002), Arenal (Hagerty et al., 2000), and Shishaldin (Petersen and McNutt, 2007).

However, three explosions (Nos 6, 7, 8 in Table IV) from 13 studied were characterized by the delay times greater the mean values of this parameter. This might indicate that these particular explosions were deeper or that a delay occurred between the onset of gas and ash emission and the emission of acoustic energy. The source depths of Vulcanian explosions at Popocatepetl between 0 and 220 m beneath the crater, obtained from the time delay between acoustic and seismic arrivals, differ from the source depths of Vulcanian explosions at the same volcano (1.5 - 1 km beneath the crater) obtained from the polarization and the inversion of VLP seismic waves of these explosions (Arciniega-Ceballos et al., 1999; Chouet et al., 2005). As was shown by Kumagai et al. (2011), the source of VLP signals of explosions may be associated with the deep ascent of magma in the conduit that later generates the explosion reaching the surface. Therefore, the source of explosions obtained for VLPs at Popocatepetl by (Arciniega-Ceballos et al., 1999; Chouet et al., 2005) will be deeper than the source of acoustic signals, generated when magma reaches the surface.

CONCLUSION

Popocatepetl has been recently generating Vulcanian explosions that were observed with acoustic and seismic data. For the seismic data, some higher magnitude explosions had their energy distributed in four different frequency bands, VLP, LP, SP and HF, each band has been modelled as representing a different process, each of which was occurring simultaneously. The calculated seismic energy was within the same range as has been published for Vulcanian explosions at other andesitic volcanoes. In the case of the acoustic energy, this was slightly larger than the seismic energy in most cases, especially for the largest explosions. Some similarity was also observed when comparing the acoustic energy with that of Vulcanian explosions at other volcanoes. The higher values of acoustic energy with respect to the seismic energy were indicated by the VASR being moderate (1-10) to large (> 10). This most likely implies a short and wide conduit and/or a small source region, a product of the Vulcanian activity due to the obstruction of the conduit by an impermeable cap or lava dome. The times of the delay between the seismic and acoustic signals, show that the depth of the acoustic-seismic signals source is shallow, less than approximately 220 m beneath the lava dome or crater floor. It follows that a pathway exists between the VLP source, located at 1.5-1 km beneath the crater, as observed in other studies, and the acoustic-seismic source.

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Chapter 3

**EXPERIENCE OF THE DETECTION OF ASH PLUME
AND ESTIMATION OF ITS HEIGHT USING LOCAL
SEISMICITY FOR KAMCHATKAN VOLCANOES
DURING 2003-2011
(KAMCHATKA PENINSULA, RUSSIA)**

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ABSTRACT

In 2003 we began using an original empirical method for detecting ash plumes and estimating plume heights using local seismicity. The method includes two stages: 1) Analysis of spectrograms from seismic events, which correspond to ash plumes. These show a predominant frequency content increasing from 1 Hz to 2-4 Hz, which is unusual for other local volcanic earthquakes. 2) Correlation of the amplitude of the fixed seismic signal envelope (absolute ground velocity) with the ascent rate of the ash plume based on continuous video observation. Fixed correlation allows us to use the integral of absolute ground velocity (or cumulative absolute ground displacement) for ash plume height estimation. Utilization of this method in real-time regime gives evidence of high performance in Kamchatka. It allows us to have fast and adequate estimates for ash plume height in absence of visual and video data. The accuracy of our method is ~30% error in height for Karymsky, Kizimen and Bezymianny volcanoes, and ~50% error in height for Sheveluch volcano. More than 680 urgent messages about real and possible dangerous ash plumes for aviation were sent within international KVERT project to Tokyo, Anchorage and Washington VAACs, Alaska Volcano Observatory, Institute of Volcanology and Seismology in Petropavlovsk-Kamchatsky and the Elizivno Airport from

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01 January 2006 to 01 May 2011, including: Sheveluch volcano ~400, Karymsky ~ 230, Bezymianny - 10, and Kizimen – 46 messages.

INTRODUCTION

Assessing volcanic hazards using remote methods, regardless of weather condition and the presence of observers, remains rather critical today, especially for flight safety (Kiryanov, 1992; Kiryanov et al., 2001; Perkins, 2003; Neal et al., 2009). Real-time visual and video observations have an undoubted priority, but few of the total number of plumes could be visually confirmed in Kamchatka due to darkness (about 50% of the time) and frequent poor weather. In addition, more than half of the active volcanoes are located far from population centers and continuous visual (or video) observations are missing for these.

Satellite observations play an important role in volcano monitoring (Roach et al., 2001). Satellite data provide quantitative information on volcanic clouds (Dean et al., 2002; Schneider et al., 2000) and elevated surface temperatures at volcanic edifices (Dehn et al., 2000; Harris et al., 1997, 2000; Wright et al., 2004). Satellite data cover large areas during night and day, but they are dependent on weather conditions. These data are intermittent and available with a delay (at least 30 minutes and 2 hours on average). Therefore it is difficult to accurately determine an eruption onset from satellite data.

Seismicity has long been one of the most commonly monitored aspects of active volcanoes because of its 24 hours per day reliability regardless of weather. Traditionally determination of the beginning of explosive eruptions is based on seismic data, and estimates of the height of ash emissions are made using the amplitude (or displacement) of seismic signals (Endo and Murray, 1991; Murray et al., 1990; McNutt, 1994; Prejean and Brodsky, 2011). Recently good results have been obtained also by infrasound observations to determine onset of ash emissions and estimates of ash plume height and volume (Fee et al., 2010; Matoza et al., 2009; Firstov, 2003). But there is no real-time infrasound monitoring of Kamchatka volcanoes. Thus seismic monitoring is the leading method in Kamchatka for tracking volcanic activity.

With sufficient experience at a given volcano during an extended eruption sequence, seismic monitoring can be used to assess activity remotely. Each volcanic event (lava flow, ash plume, avalanche etc.) produces its own special seismic signal. Spectral analysis allows us to divide these seismic signals into many types.

Uninterrupted, time-stamped video observation allows us to interpret and measure the absolute dimensions of some volcanic events. It is possible to create a database of volcanic event descriptions and their corresponding seismic signals, enabling one to infer an ash plume or avalanche event solely on the basis of seismicity (Kozhevnikova, 2008).

Previous works, which investigated the correlation of seismic records and ash plumes, had several problems due to approximate or incomplete visual data (McNutt, 1994):

- 1) Very often, inexperienced people or non-technical observers (pilots, tourists) reported explosive eruptions, while this "eruption" was in fact only a gas-steam emission.
- 2) For all practical purposes, it is often impossible to differentiate an ash plume from gas steam plume, especially in twilight (sunset or sunrise) even for experienced

observers. During twilight (or when backlit), a gas-steam plume is often darker than during the more morning or afternoon.

- 3) Precise times are not included in some observations.
- 4) Plume height estimates may have large error due to a lack of a calibrated reference.

In Kamchatka a good opportunity to calibrate ash explosions from seismic data appeared beginning from 9 October 2000 when the first video camera of the Kamchatkan Branch of Geophysical Survey (KBGS) RAS was installed at seismic station "Kluchi" to monitor Klyuchevskoy Volcano. Similar video systems were installed on 17 May 2002 to observe Sheveluch Volcano, on 20 August 2003 to observe Bezymianny Volcano, and on 25 July 2011 to observe Kizimen Volcano. All continuous video systems have an image timing precision of ± 1 sec (GPS correction). Information is available online: <http://www.emsd.ru/>.

METHOD

This Chapter presents our original empirical (not theoretical) method of remote determination of ash plume initiation and height using local seismicity. This method was first tested for Sheveluch volcano during a period of ongoing eruption and generation of ash clouds by processes of explosive lava dome collapse or explosive tephra emission from the growing dome and was based on two experimentally established facts (Senyukov et al., 2004):

- 1) All visually confirmed (by observer or video) ash plumes were accompanied by seismic signals of similar character. Spectrograms for these events show a peak frequency increasing over the course of the event from 1 Hz to 2-4 Hz, which is unusual for other volcanic earthquakes. Strong ash plumes accompanied by pyroclastic flows had further peak frequencies increasing from 3 Hz to 5 Hz towards the end of the signal. Similar increases in peak frequency were also noted during seismic observations at other actively erupting volcanoes (Chouet et al., 1997, 2003; Jellinek et al., 2011; Ripepe et al., 2001; Zobin et al., 2006; Zobin, 2012). To explain the origin of the low- and high-frequency impulses, Ripepe et al. (2001) carried out laboratory experiments and inferred that low-frequency seismic signals are generated by the rapid expansion of gas in a magma conduit, whereas high-frequency signals are generated by explosions at the magma's free surface.
- 2) Due to continuous video observations, we were able to correlate the amplitude of the seismic signal envelope (absolute ground velocity) with the ascent rate of the ash plume. This correlation allowed us to use the integral of absolute ground velocity (or cumulative absolute ground displacement) for the ash plume height estimation.

We apply this method in real-time regime to four active volcanoes in Kamchatka: Sheveluch, Karymsky, Bezymianny and Kizimen (Figure 1, Table 1). A map of Kamchatka is shown in Figure 1 with the research areas marked as squares. The seismic stations are shown with the white triangle symbols. All seismic stations are equipped with 3-component short-period seismometers with an eigenfrequency of 1.2 Hz. Seismic registration of velocity

allows us to record the seismic signals within 0.5 - 20 Hz spectral range and ~ 54 dB dynamic range, for a velocity limit value of about ~ 40 microns per second. Unfortunately, station “SVL” was destroyed during a major eruption on February 28, 2005. This station is shown with the black triangle symbol in the Figure 1.

DATA

Different volcanic processes produce different types of earthquakes, varying in waveform duration, onset, frequency content, and amplitude.

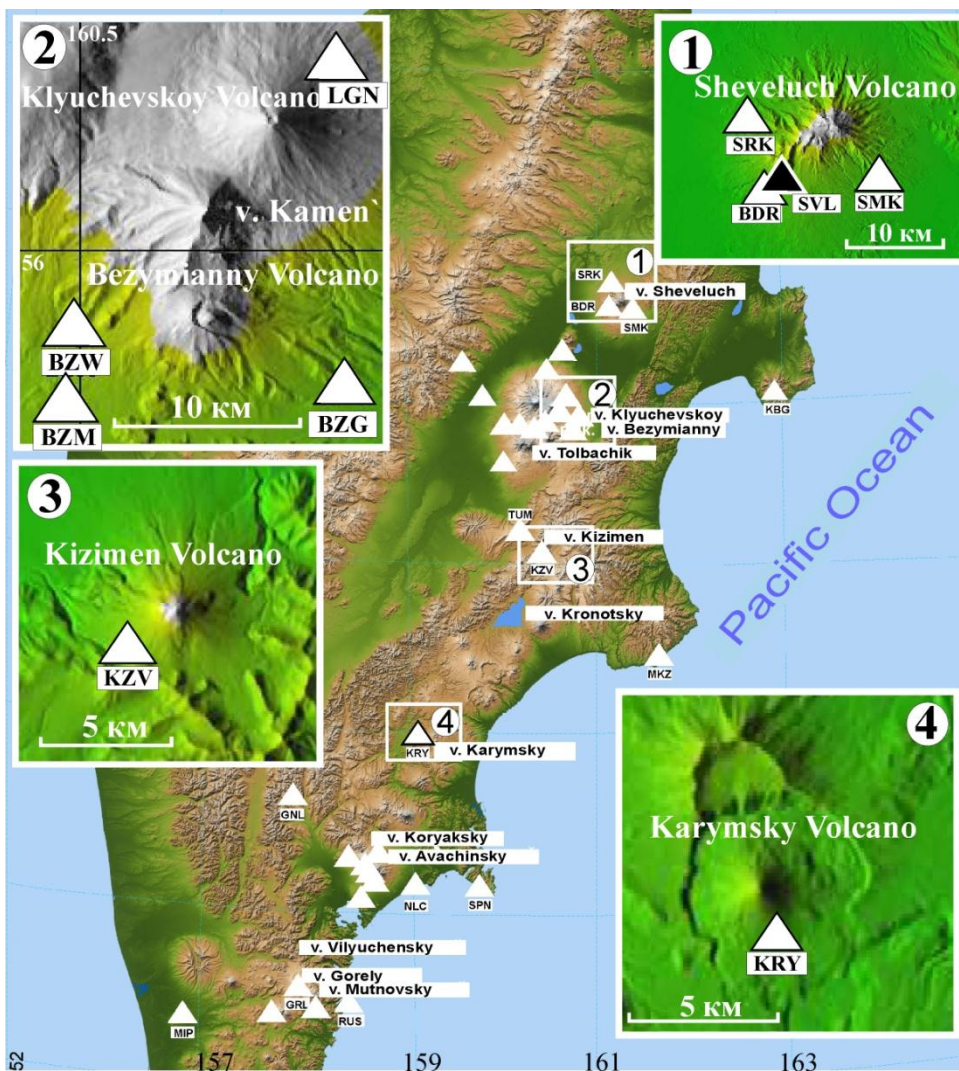


Figure 1. Map of Kamchatka Peninsula. Four volcanic areas are marked on the map: 1) Sheveluch volcano; 2) Bezymianny volcano; 3) Kizimen volcano; 4) Karymsky volcano. Seismic stations are shown with the white triangle symbol.

Table 1. Volcano data and distance of the seismic station from volcanic ash source

Name	Coordinates		Elevation, m	Code of seismic station and distance (D) from the ash source, km
	Lat., N	Long., E		
Sheveluch (active dome)	56.63	161.32	2600	SVL(D=8.5), SRK(D=9.9), SMK(D=10.4), BDR(D=10.2)
Bezymianny	55.97	160.60	2880	LGN(D=13.6), BZW(D=6.4)
Kizimen	55.13	160.33	2375	KZV(D=2.6)
Karymsky	54.05	159.44	1500	KRY(D=1.4)

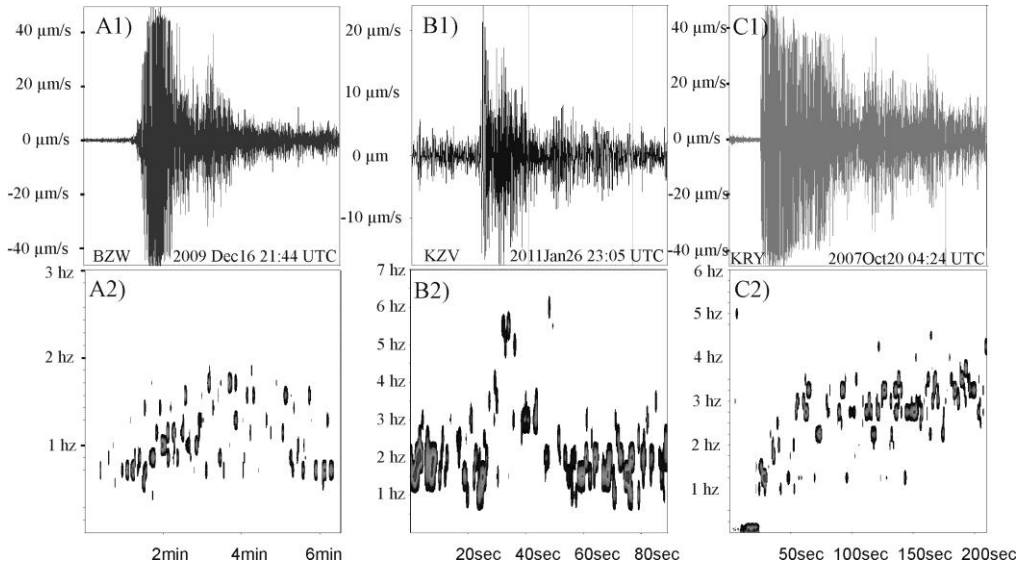


Figure 2. A1) the seismic signal associated with Bezymianny’s ash plume (16 December 2009, height ~8 km above sea level) and A2) its spectrogram; B1) the seismic signal associated with Kizimen’s ash plume (26 January 2011, Figure 4B) and B2) its spectrogram; C1) the seismic signal associated with Karymsky’s ash plume (20 October 2007, Figure 4C) and C2) its spectrogram.

We selected and classified hundreds of waveforms and the associated volcanic process using photos and video data. The resulting database of event standards allows us to make inferences about the volcanic event types according to seismic data only. The examples of seismic signals and their spectrograms, associated with the appearance of ash plumes at Bezymianny, Kizimen and Karymsky volcanoes, are presented in Figure 4.

The program DIMAS (Droznin and Droznina, 2010) was used for the signal processing, including analysis in the time and frequency domain, spectrogram calculation, and computation the envelope and integral of the signal. Very often we investigated clipped seismic signals (~1000), but we are sure that clipping of signal is not important to calculate spectrogram (Figure 3). Figure 3 presents unclipped seismic signal (B1) of ash plume (30 October, 2004, 03:04 UTC, height is 1500 m above the dome) and its spectrogram (C1). This seismic signal was cut to 50% (Figure 3B2) and to 25% (Figure 3B3) levels from the limit value. Despite this processing, the spectrograms of these cut signals remain essentially unchanged (Figure 3, C2 and C3). Figure 3 (D) presents the power spectra of the seismic signals from Figure 3 (B1, B2 and B3).

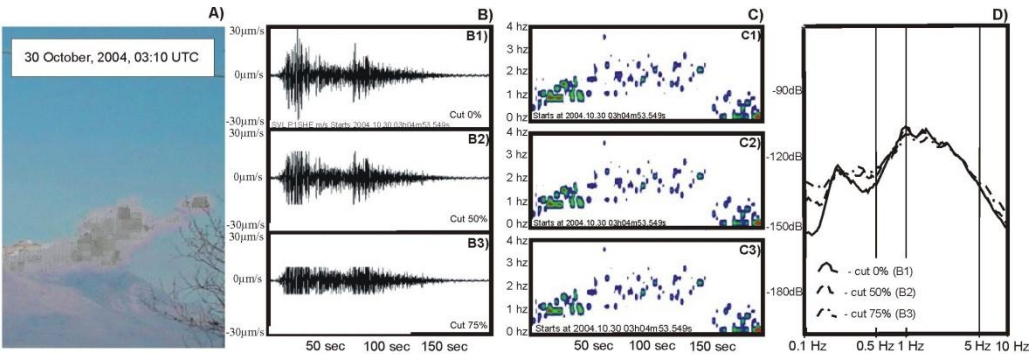


Figure 3. A) Sheveluch’s ash plume, 30 October 2004 at 03:04 UTC, height is 1500 m above the dome; B) seismic signals: B1) real seismic signal registered by seismic station “SVL”, B2) real signal was cut to 50 % level from maximum possible value, B3) real signal was cut to 25 % level from maximum possible value; C) spectrograms: C1) spectrogram of signal B1, C2) spectrogram of signal B2, C3) spectrogram of signal B3; D) power spectra of the signals B1, B2 and B3.

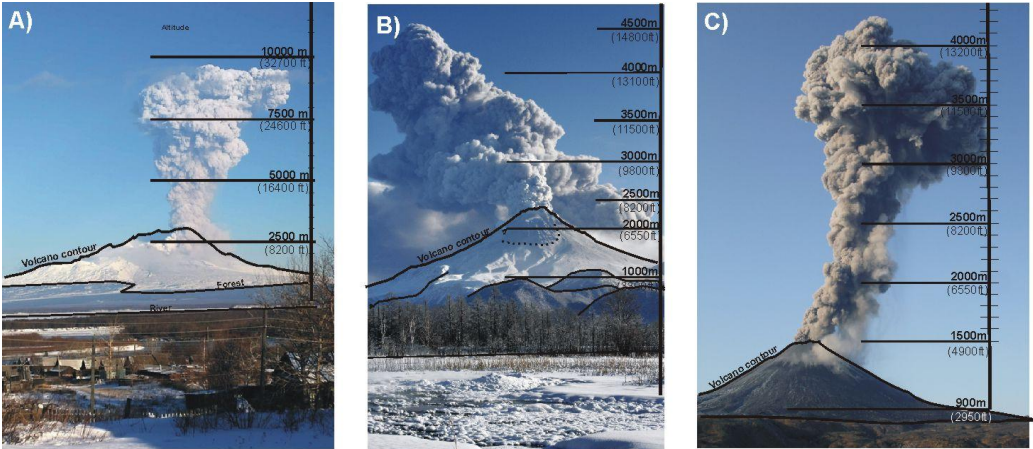


Figure 4. Photos with the reference profiles of volcanoes and the height scale above sea level: A) Sheveluch’s ash plume, 17 December 2006, 01:58 UTC, taken by Yury Demyanchuk; B) Kizimen’s ash plume, 26 January 2011, 23:09 UTC, taken by Alexandr Sokorenko; C) Karymsky’s ash plume, 20 October 2007 04:26 UTC, taken by Valery Yudin.

All three lines are similar within the investigated frequency range from 0.5 to 5 Hz. Consequently, our technique of spectral analysis can correctly detect ash plumes even when the seismic signal is clipped.

To measure the height of an ash plumes, we used the photo and video data only. Images with variable focus distances were reconciled using the CorelDraw program. The profiles of volcanoes were scaled using well-known altitudes of prominent features points (Figure 4). In the future, this profile will be not changed; therefore it provides a stable reference point for measuring distances. Each photo or video image of the volcano was resized to match the reference volcano profile, and each ash plume could be measured by a consistent altitude scale, improving our precision of height measurements. The linear scale of Figure 4 is not correct due to the 3-dimensionality of the field of view.

However, calculated errors of nonlinear scale (2% at 6 km elevation, 3% at 10 km elevation and 5% at 15 km elevation) are considerably less than possible errors due to deviation of ash column from vertical line due to winds aloft.

Figure 5 presents the comparison of the seismic signal and its spectrogram recorded during an explosion of Sheveluch volcano on 16 September 2002 at 23:58 UTC with the variations in the height of ash plume. All graphs have a similar horizontal time scale. The ash plume appeared at the earth surface in 15 second after the corresponding seismic signal. The peak frequency of this signal was $\sim 1\text{--}1.5$ Hz during the first 20 seconds. Later, the peak frequency increased up to 2 Hz for 50 seconds and up to 3 Hz for 3 minutes.

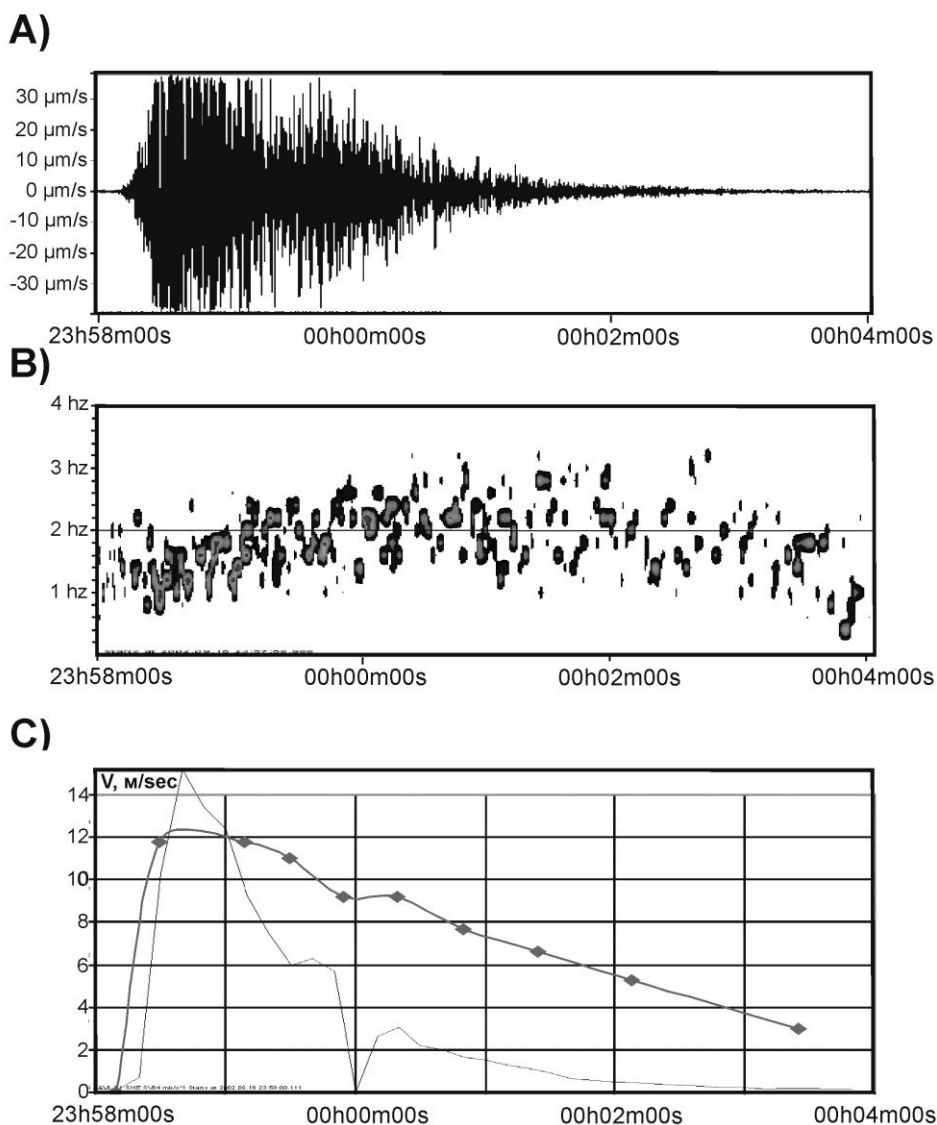


Figure 5. Seismicity and derived rise velocity history for Sheveluch's ash plume, 16 September 2002 at 23:58 UTC: A) seismic signal recorded by seismic station "SVL"; B) spectrogram; C) comparison of the seismic signal envelope (dimensionless units) (shown with thin line) with the variations in the ash plume rise velocity (vertical scale) (shown with thick line with points).

This peculiarity was noted only for the seismic signals accompanied by the ash plumes. Figure 5 (C) demonstrates a correlation of the seismic signal envelope and the ash plume rise velocity. The graph of the rise velocity was obtained by movie investigation using the altitude scale in Figure 4(A). To obtain a graph of the seismic signal envelope, we used the average value of the absolute ground velocity within the 10-second windows. The ash cloud height is an integral of the rise velocity function with time, so it is a logical result to substitute a function of velocity with a function of the seismic signal envelope (or absolute ground velocity) using experimentally determined coefficients.

We offer a possible interpretation of Figure 5. First of all, it is important to note that the activity, documented in Figure 5, represents the explosive eruption of a two-phase mixture of gas and ash, unlike the gas-steam emissions usually characterizing the lava dome activity. As it is supposed by (Sahagian, 1999; Gardner et al., 1996), due to the rapid fragmentation of magma in the conduit, the magma reaches the surface, where produces an explosion and generates the ash column. The process of generation of the ash column produces a seismic signal with relatively high frequencies, more than 2 Hz for Sheveluch volcano, according to our observations. We suppose the following increase of the peak frequency, up to 4 Hz, as a result of the avalanches triggered by the fall of ash and bomb. We interpret the 15-second delay between the onset of the seismic signal and appearance of the ash plume as the interval between the initial gas burst and opening of a pathway to the surface and the arrival of the gas-ash mixture at the surface. A similar delay between onset of seismic signal and appearance of the ash plume was found at Volcán de Colima (Zobin, 2012).

APPLICATION OF THE METHOD FOR VOLCANO MONITORING

Sheveluch Volcano

Sheveluch is the northernmost active volcano of the Kamchatka Peninsula (Figure 1). It is one of the largest volcanic structures in Kamchatka. The present structure includes three main units: Stary (Old) Sheveluch, the old caldera, and the historically active crater - Molodoy Sheveluch. The volcano is situated in the northern part of the central Kamchatka depression, 45 km northeast of Kluchi town. During historic times (XIX-XX centuries), Sheveluch has had two types of eruptions: (a) catastrophic explosions with accompanying directed blasts (1-2 March 1854 and 12 November 1964); (b) weak or moderate eruptions that accompanied the growth of extrusive dacite domes (Melekestsev et al., 1991). Correlations between the height of Sheveluch's ash plumes and the integral of absolute ground velocity at station "SVL" are shown in Figure 6. This station was located 8.5 km to the southwest from the active dome. We analyzed 214 events confirmed by visual, photo and video observations between 1999 and 2004. These plumes had the following heights: minimum=200 m, maximum=15000 m, average height=2142 m.

The data shown in Figure 6 are correlated with the coefficient of correlation $R = 0.82$; mean-square error for the height estimation is 1018 m, or 48% of average height. The coefficient of determination $R^2 = 0.67$, shown in Figure 6, indicates that 67% of the total variation in the plume height can be explained by the linear relationship between the integral of absolute ground velocity and the plume height.

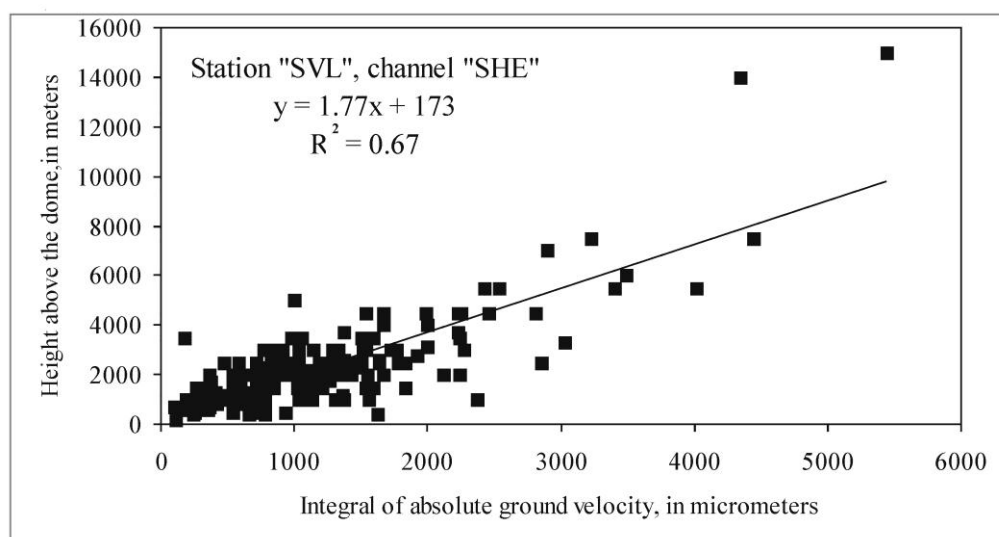


Figure 6. Relationship between the height of Sheveluch's ash plumes and the integral of absolute ground velocity at station SVL (Figure 1).

The method was tested in real-time during 2003-2004. A total of 866 possible ash explosions were identified with seismic data. Video observations confirmed 72 of seismically identified explosions, or ~8%. The remained seismic signals were recorded when the volcano was closed by clouds or during the dark time.

Unfortunately, station Sheveluch "SVL" was destroyed during a major eruption on 28 February 2005. New radio telemetered seismic stations were installed on the slopes of Sheveluch volcano in the autumn of 2005 (Figure 1). New stations are located in safer places, but the signals from ash eruptions are not recorded as clearly as it was before. However, we determined new coefficients to calculate the height of ash plumes from integral of absolute ground velocity using new data, and our work in real-time continues. Correlations between the height of Sheveluch's ash plumes and integral of absolute ground velocity for three new stations are shown in Figure 7.

We analyzed 150 ash-plume events confirmed by visual, photo and video observation between 2005 and 2011. These plumes had the following heights: minimum=600 m, maximum=10000 m, average height=2684 m. We calculated the parameters fitting the data in Figure 7: average correlation coefficient for three stations $R=0.51$, the average mean-square error is 1227 m, or 46% of the average height. A coefficient of determination R^2 ranged from 0.23 to 0.29 indicating low linear relationship between the integral of absolute ground velocity and the plume height (Figure 7).

Karymsky Volcano

Karymsky volcano is one of the most active volcanoes in Kamchatka and is located in the central part of the Eastern volcanic belt, roughly 120 km north-east of Petropavlovsk-Kamchatcky (Ivanov et al., 1991).

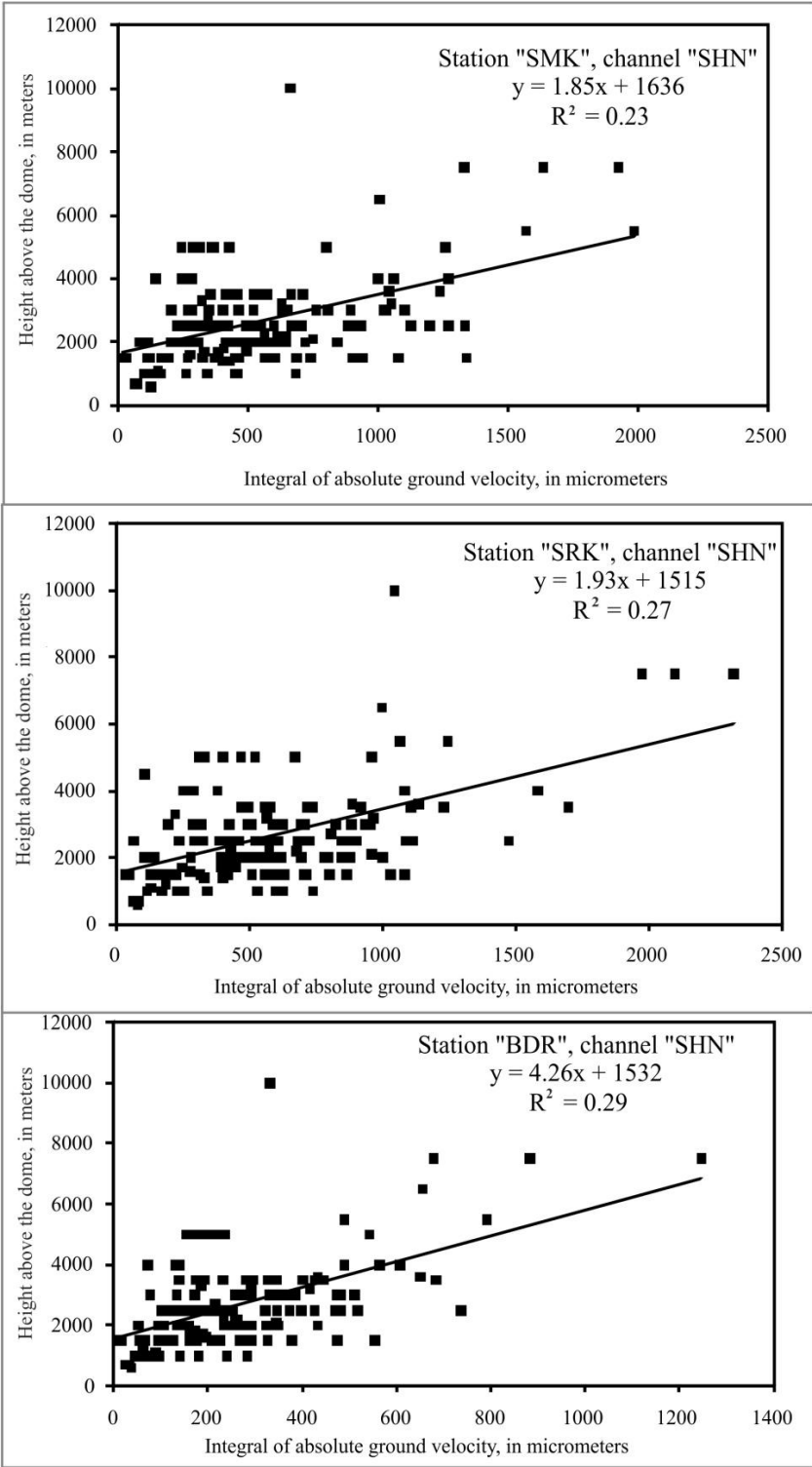


Figure 7. The relationships between the height of Sheveluch`'s ash plumes and the integral of absolute ground velocity at stations SMK, SRK, BDR (Figure 1).

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The most recent eruption of Karymsky volcano began in 1996 and has continued to the present with the short periods of quiescence. Karymsky is removed from human settlements, so there is no possibility of carrying out continuous visual observations.

We compiled information (photo, visual and video) of various volcanic events using field investigation of the Institute of Volcanology and Seismology and compared these observations with the accompanying seismic signals (Kozhevnikova, 2008). The Kamchatkan Branch of Geophysical Survey has a single seismic station KRY, which is located on the southeast slope of Karymsky 1.4 km from crater (Figure 1).

Figures 8 and 9 show the relationships between the height of ash plumes at Karymsky volcano and the integral of absolute ground velocity obtained for two intervals of observations with different seismic instruments.

At first we analyzed 70 events confirmed by visual, photo and video observation between 2004 and 2007 together with the seismic data obtained with the short-period SHE channel (Figure 8). The ash plumes during this period had the following heights: minimum=200 m, maximum=5000 m, average height=1703 m.

The relationship between the height of ash plumes and the integral of absolute ground velocity is characterized by the correlation coefficient $R=0.73$ ($R^2=0.53$) (Figure 8); the mean-square error is 834 m, or 45% of average height. The channel SLZ (a short-period channel of low sensitivity) was installed on the station KRY in September 2006 that allowed registering of stronger seismic events without limitation of the signal amplitude. Figure 9 shows the relationship between the height of Karymsky's ash plumes and the integral of absolute ground velocity for 38 events confirmed by visual, photo and video observations between 2007 and 2011. These plumes had the following heights: minimum=200 m, maximum=3000 m, average height=1418 m. The correlation coefficient, obtained for these data ($R=0.84$), and corresponding coefficient of determination ($R^2=0.71$) (Figure 9) were higher than for the clipped seismic signals obtained with SHE channel.

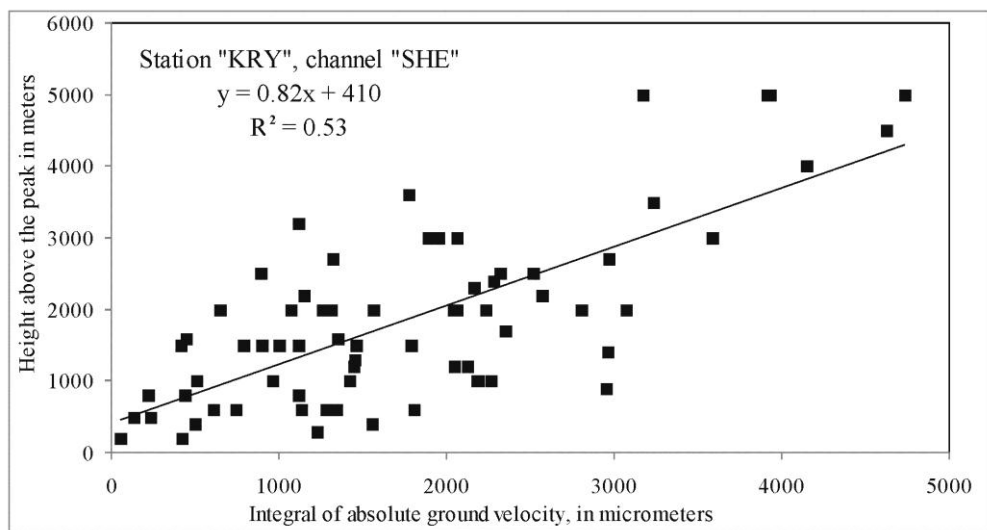


Figure 8. Relationship between the height of Karymsky's ash plumes and the integral of absolute ground velocity at station KRY (channel SHE) (Figure 1).

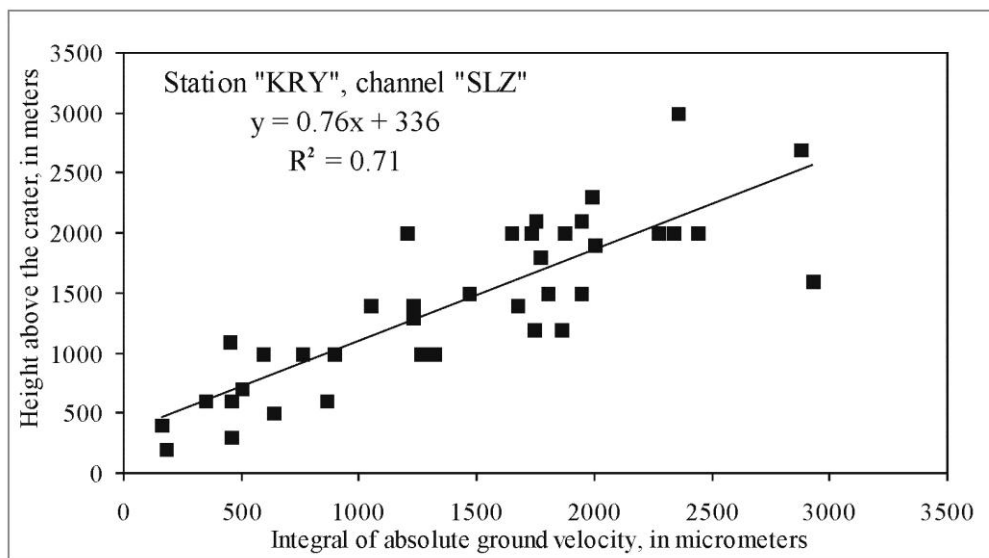


Figure 9. Relationship between the height of Karymsky's ash plumes and the integral of absolute ground velocity at station KRY, channel "SLZ" (Figure 1).

So using nonclipped signals improved the correlation coefficient and decreased the mean-square error equal now to 358 m (or 25% of the average height).

Kizimen Volcano

Kizimen is the southernmost of the active volcanoes of the Central Kamchatka Depression (Shantser et al., 1991). The latest eruption began at the end of 2010 after ~80 years of quiescence, and was ongoing in July 2012.

The eruption was preceded by strong and long-time seismicity starting in April 2009 (Senyukov et al., 2011). The early phase of the eruption was characterized by new and increased fumarolic activity beginning from 16 October 2011 and episodic ash emissions beginning from 11 November 2011 (Senyukov et al., 2011).

According to the seismic (Senyukov et al., 2011), satellite (Melnikov et al., 2011) and visual data of the Kronotsky Biosphere Nature Reserve, the explosive events with 6-10 km height above sea level and up to 3000 km extension occurred on 12 and 31 December 2010, 6, 12 and 15 January, 7 March, and 2 May 2011. We used information (photo, visual and video) obtained from the Institute of Volcanology and Seismology and the Kronotsky Biosphere Nature Reserve. We analyzed seismic signals that were recorded by the station KZY located on the southeast slope, 2.6 km from the summit peak. The correlation between the height of Kizimen's ash plumes and the integral of absolute ground velocity at station KZY, channel SHN, are shown in Figure 10. We analyzed 19 events confirmed by visual, photo and video observation in 2011. These plumes had the following heights: minimum=1000 m, maximum=4000 m, average height=2236 m. We calculated the following parameters for the data plotted in Figure 10: correlation coefficient $R=0.74$, coefficient of determination $R^2=0.55$, mean-square error was 594 m, or 27% from average height.

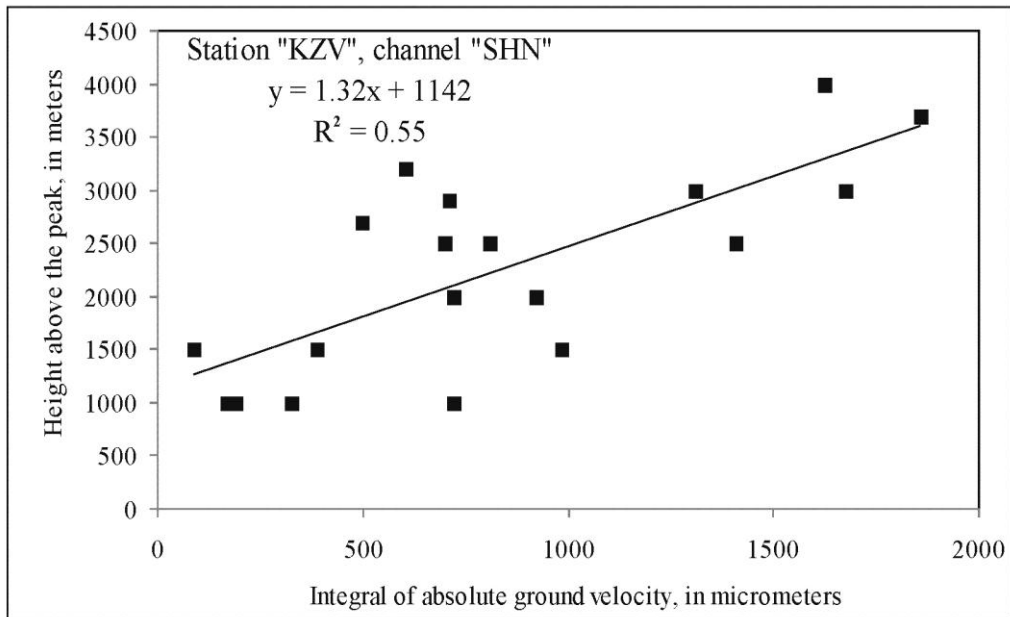


Figure 10. Relationship between the height of Kizimen's ash plumes and the integral of absolute ground velocity at station KZY (Figure 1).

Bezmianny Volcano

Bezmianny volcano is located in the central part of the Klyuchevskaya group of volcanoes southwestward of Klyuchevskoy volcano (Bogoyavlenskaya et al., 1991). The last catastrophic eruption occurred on 30 March 1956, after 900-1000 years of quiescence. Since 1956 the volcano had 1-2 explosive eruptions per year. The heights of the ash plumes ranged from 6 km to 15 km above sea level.

We analyzed 9 events confirmed by visual, photo and video observation between 2000 and 2011 (Figure 11). These plumes had the following heights: minimum=3000 m, maximum=10000 m, average height=5722 m. We calculated the parameters for data plot in Figure 11: correlation coefficient $R=0.73$, coefficient of determination $R^2=0.54$, mean-square error is 1472 m, or 26% from average height.

ILLUSTRATION OF THE REAL-TIME VOLCANO MONITORING

We began to use this method in real-time monitoring in 2003. To detect seismic signals, corresponding to ash plumes, we used a special program (author – Dmitry Droznin), which processed the seismic signals every hour during working hours (from 08:00 a.m. till 06:00 p.m. local time each day).

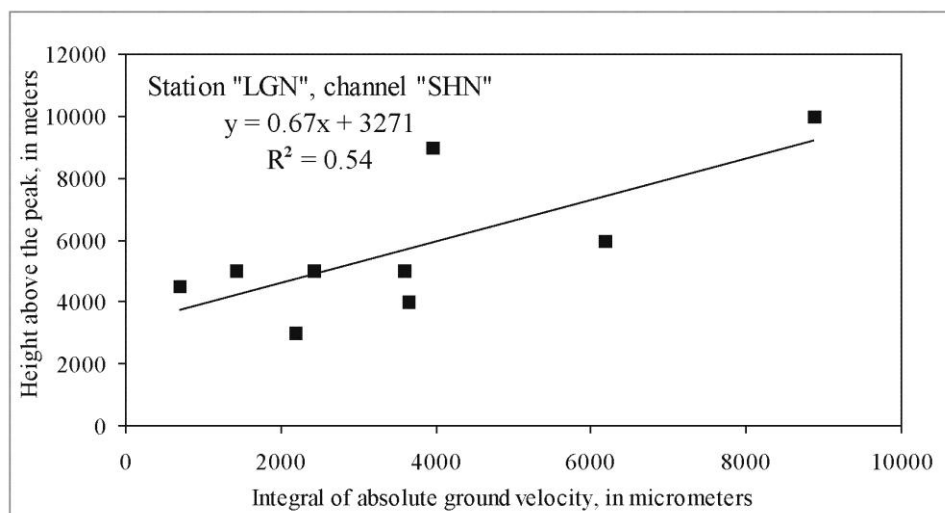


Figure 11. Relationship between the height of Bezymianny's ash plumes and the integral of absolute ground velocity at station LGN (Figure 1).

This program calculated spectrograms. When a possible ash plume was detected seismically, we collected all video, visual and satellite data about this ash plume and published this information on the web site of KBGS (<http://www.emsd.ru/~ssl/monitoring/main.htm>). More than 680 urgent messages about real (if there was video or visual confirmation) or possible (if ash plume was detected using seismic data only) dangerous ash plumes (mostly with the height >6 km above sea level) for aviation were sent, in the frame of the international KVERT project, to Tokyo, Anchorage and Washington Volcanic Ash Advisory Centers, the Alaska Volcano Observatory, the Institute of Volcanology and Seismology (KVERT group) in Petropavlovsk-Kamchatsky and the Elizivo Airport from 1 January 2006 to 1 May 2011, including: Sheveluch volcano ~400, Karymsky ~230, Bezymianny - 10, and Kizimen - 46 messages.

Since June 2012 our laboratory began to use the automatic program (author - Vitaly Bliznezov) to detect ash plumes from seismic data, which run every 15 minutes. As an example of our work we show a protocol of ash plume registration on 27 July 2012:

- 1) 01:58 UTC – beginning of seismic signal of possible ash plume at Sheveluch volcano;
- 2) 02:15 UTC – automatic program detected possible ash plume and sent this information via email to laboratory staff;
- 3) 02:32 UTC –scientist-on-duty checked the seismic record and its spectrogram, calculated height of possible ash plume and sent the following urgent information via email according to regulations:Information from KB GS: Sheveluch (1000-27) 01:58 UTC on July 27, 2012 Volcano:
Sheveluch (1000-27)Location: 56°39'N, 161°21'E Elevation of vent (meters/feet): 2,500 m/8202 ft the dome elevation Height of ash plume (km/feet) ASL and how determined: > 10 km /32800 ft - seismic dataDistance from the volcano (km/mi): unknown Direction of ash plume or ash cloud drift from the volcano: unknown Start time of explosion and how determined: 01:58 UTC -seismic data Duration of

eruption (or indicate eruption is continuing): eruption is continuing Comment: According to seismic data, possible ash plume and avalanche occurred on July 27 at 01:58 UTC. If an ash plume was generated, the approximate height of this ash plume more to 10000m (~32800ft) ASL. Video data: obscured by clouds.

- 4) 02:47 UTC, Tokyo VAAC sent “VA (volcanic ash) ADVISORY” about this plume, although “VA NOT IDENTIFIABLE FROM SATELLITE”, with remark “height of ash estimated by seismic record”, <http://ds.data.jma.go.jp/svd/vaac/data/index.html>;
- 5) 04:17 UTC, Tokyo VAAC sent “VA ADVISORY” about ash plume extension to northeast according to satellite data.

This example shows a special and difficult case, when video and satellite observations were practically impossible due to a bad weather. We found additional confirmation of this ash plume on the web site of Institute of Volcanology and Seismology (Petropavlovsk-Kamchatsky) on the next day only. Video image captured a strong ash plume at 02:20 UTC on 27 July when visual observation of Sheveluch was possible for a very short time only in a small window between clouds. Thus we confirmed a strong ash plume extending to northeast.

All observation methods have limitations and errors. We think only integration of all possible data in real time monitoring allows to detect volcanic ash plumes more accurately.

CONCLUSION

A technique for estimation of ash cloud height using seismic data only was designed and implemented in monitoring activity at four active Kamchatkan volcanoes. This method offers a more reliable method to warn of ash clouds that can affect air traffic, especially during bad weather or when visual observations and confirmation are not possible. The accuracy of our method for now is ~30% error in height for Karymsky, Kizimen and Bezymianny volcanoes, and ~50% error in height for Sheveluch volcano. From our experience, we can assume that the installation of broadband stations with large dynamic range at a distance not more than 5 km from the active volcanic centers will allow more efficient and correct use of this method. It should also be noted that wind speed has a big effect on the height of ash emission, but now we cannot use this information in our estimation.

The using of this method for Klyuchevskoy’s ash plumes (Figure 1) gave poor results. Klyuchevskoy volcano is the most active and powerful basaltic volcano of the Kuril-Kamchatka volcanic area (Khrenov et al., 1991). The summit eruptions of this volcano were preceded and accompanied by strong volcanic tremor. We fixed many strong ash emissions with height up to 5 km above the crater (or up to 10 km above sea level) during the study period, but we could not detect seismic signal corresponding ash plume using spectral analysis. Thus we think our method will not be useful for basaltic volcanoes.

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Chapter 4

VOLCANO MONITORING AND EARLY WARNING ON MT. ETNA, SICILY BASED ON VOLCANIC TREMOR: METHODS AND TECHNICAL ASPECTS

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ABSTRACT

Eighteen paroxysmal episodes occurred on Mt Etna in 2011, and provided rich material for testing automatic procedures of data processing and alert systems in the context of volcano monitoring. The 2011 episodes represent a typical picture of activity of Mt Etna: in 2000 and 2001, before the 2001 flank eruption, more than one hundred lava fountains were encountered. Other major lava fountains occurred before the flank eruptions of 2002/03 and 2008. All these fountains, which are powerful but usually short lived phenomena, originated from the South-East Crater area and caused the formation of thick ash clouds, followed by the fallout of material with severe problems for the infrastructure of the metropolitan area of Catania.

We focus on the seismic background radiation – volcanic tremor – which plays a key role in the surveillance of Mt Etna. Since 2006 a multi-station alert system has been established in the INGV operative centre of Catania exploiting STA/LTA ratios. Besides, it has been demonstrated that also the spectral characteristics of the signal changes correspondingly to the type of volcanic activity. The simultaneous application of Self Organizing Maps and Fuzzy Clustering offers an efficient way to visualize signal characteristics and its development with time, allowing to identify early stages of

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eruptive events and automatically flag a critical status before this becomes evident in conventional monitoring techniques.

Changes of tremor characteristics are related to the position of the source of the signal. The location of the sources exploits the distribution of the amplitudes across the seismic network. The locations were extremely useful for warning throughout both a flank eruption in 2008 as well as the 2011 lava fountains, during which a clear migration of tremor sources towards the eruptive centres could be noticed in advance. The location of the sources completes the picture of an imminent volcanic unrest and corroborates early warnings flagged by the changes of signal characteristics.

On-line data processing requires computational efficiency, robustness of the methods and reliability of data acquisition. The amplitude based multi-station approach offers a reasonable stability as it is not sensitive to the failure of single stations. The single station approach, based on our unsupervised classification techniques, is cost-effective with respect to logistic efforts, as only one or few key stations are necessary. Both systems have proven to be robust with respect to disturbances (undesired transients like earthquakes, noise, short gaps in the continuous data flow), and false alarms were not encountered so far.

Another critical aspect is the reliability of data storage and access. A hardware cluster architecture has been proposed for failover protection, including a Storage Area Network system. We outline concepts of the software architectures which allow easy data access following predefined user policies. We envisage the integration of seismic data and those originating from other scientific fields (such as volcano imagery, geochemistry, deformation, gravity, magneto-telluric), in order to facilitate cross-checking of the findings encountered from the single data streams, in particular allowing their immediate verification with respect to ground truth.

1. INTRODUCTION

Volcanic areas often offer rich natural resources, such as fertile grounds, or water supply. Thus many of them are situated closely to densely populated regions, forming a considerable threat to lives as well as infrastructure. The research and surveillance of phenomena related to volcanic activity has therefore become a prior interest both from a scientific point of view as well as from the needs of Civil Protection. This chapter deals with Mt Etna, Europe's largest and most active volcano. It is a typical basaltic strato-volcano formed by the superimposition of lavas and tephra deposits. Apart from lava emissions, its activity encompasses a variety of eruptive phenomena, such as spectacular lava fountains, Strombolian explosions and phreatomagmatic events.

Mt Etna is dominating the Metropolitan Area of Catania, with a population of over 1 million inhabitants and the presence of important infrastructure, such as the Intl. Airport Fontanarossa, the freeways Siracusa-Catania-Messina and Catania-Palermo, etc. The Metropolitan Area of Catania has a long history. The name of Catania appears in Greek documents as long as 2700 years ago, all this time the presence of the volcano played a key role in the daily life of people. Research on Mt Etna made bold steps ahead with the work of W. S. v. Waltershausen in the 19th century and G. Ponte, who in 1919 became director of the first University Institute of Volcanology worldwide. In the late 60's the famous volcanologist A. Rittmann founded the International Institute of Volcanology (IIV), which – together with the University Catania – became a key institution for both research and monitoring of Mt

Etna's activity. Now this activity is concentrated at the Osservatorio Etneo (INGV-OE), which is a department of the Istituto Nazionale di Geofisica e Vulcanologia.

Mt Etna's activity has been documented since Greek and Roman times. This volcano is certainly among the ones with the longest historical record of activity worldwide. Its eruptive history is marked by a large number of both flank and summit eruptions (Branca and Del Carlo, 2004). Some of the flank eruptive episodes (e.g., 1669 and 1928 eruptions) caused heavy damages in the invaded areas, strongly affecting everyday life of the population and modifying economy as well as land use. In more recent times important eruptive events occurred in the 1980s (1983, 1989), and in the 1990s (in particular 1991-93). The first decade of the present century saw a further enhanced activity with important unrests in 2001, 2002-03, 2004-05, 2006, 2007-08. Finally, eighteen lava fountains occurred in the first 11 month of 2011. All these episodes of unrest underscore the fact, that volcanic activity is continuing to touch on daily life in the Metropolitan Area of Catania and therefore is a first class problem to cope with.

The surveillance of active volcanoes poses specific problems, which are mainly logistic. Active volcanoes are obviously dangerous places, which renders the visual observation in phases of enhanced activity difficult or impossible. At the same time volcanoes form often high mountains, with adverse meteorological conditions such as low temperature, presence of snow and ice, and high wind speed. Besides, the presence of clouds poses severe limits to the use of visual or infrared camera for surveillance purposes. Therefore, in the last decades the acquisition of geophysical parameters, such as ground deformation, gravity, electro-magnetic and seismic data has experienced a rapid development. Among these, seismic data analysis has gained a key role. Seismic signals propagate effectively over long distances, permitting to place the equipment at relatively safe places, which are also easy to reach. Being modern seismic sensors extremely sensitive they allow to record even small, and short-lived phenomena in the far field.

A wide variety of seismic signals occurs on a basaltic strato-volcano like Etna. Transient signals, such as explosion quakes along with long and very long period events, are directly linked to the volcanic activity (e.g., Di Grazia et al., 2009). Earthquakes often herald and accompany the opening of eruptive fractures at the onset of flank eruptions. However, summit eruptions at Mt Etna – which are commonly more violently explosive and occur more frequently than flank eruptions – lack the occurrence of earthquakes as precursors. On the other hand, since the 1970s it has been shown (Schick and Riuscetti, 1973) that lava fountains as well as flank eruptions are commonly preceded by an increase in amplitude of the background signal, the so-called volcanic tremor, which often can be identified several hours prior to the onset of eruptive activity (see Figure 1). Monitoring of volcanic tremor has therefore become decisive for the surveillance of Mt Etna. Changes in the state of activity as well as in the eruptive style of the volcano are mirrored clearly in the signature of this signal, both in amplitude as well as in the spectral content (e.g., Alparone et al., 2003; Falsaperla et al., 2005; Langer et al., 2009).

Here we focus on three specific aspects of volcanic tremor, i.e. (i) the development of RMS-amplitudes with time, (ii) their distribution across the seismic network covering the volcanic edifice, (iii) the spectral characteristics of the signal. The RMS-amplitudes and their spatial distribution has been exploited to establish robust criteria for early warning purposes, limiting at the same time the problem of false alarms caused by the occurrence of teleseismic events, wind gusts or other noise. The same information has allowed to locate the sources of

volcanic tremor. Location and velocity of source migration in the fore-field of a volcanic unrest have repeatedly corroborated the early warning issued using threshold criteria. In a recent application of pattern recognition techniques based on spectral characteristics, it was possible to bring out further evidences for the menace of a volcanic unrest at a very early stage.

The synopsis of all these parameters concerning volcanic tremor – coincidence of amplitude increase across the network, changes in signal characteristics as well as tremor source location – allow to issue alerts at a very early stage of an unrest. The eighteen lava fountain events that occurred in 2011 have provided a sound data set for both testing the stability of the criteria as well as for their further tuning. The success, however, depends critically on the reliability and velocity of data acquisition and transmission, which bring along a considerable technological effort. The rapid accumulation of large data volumes requires the development of efficient and safe procedures of data management, both with respect to storage as well as access. The dynamics of the volcano requires that criteria used for warning purposes undergo a continuous verification and update. Further analysis of increasing time windows can be envisaged as well as the integration of seismic data with those coming from other disciplines, such as geophysical and numerical as well as volcanological, in particular in the form of imagery of surface processes. Specific data base architecture is under construction aiming at an efficient handling of large and multidisciplinary data volumes.

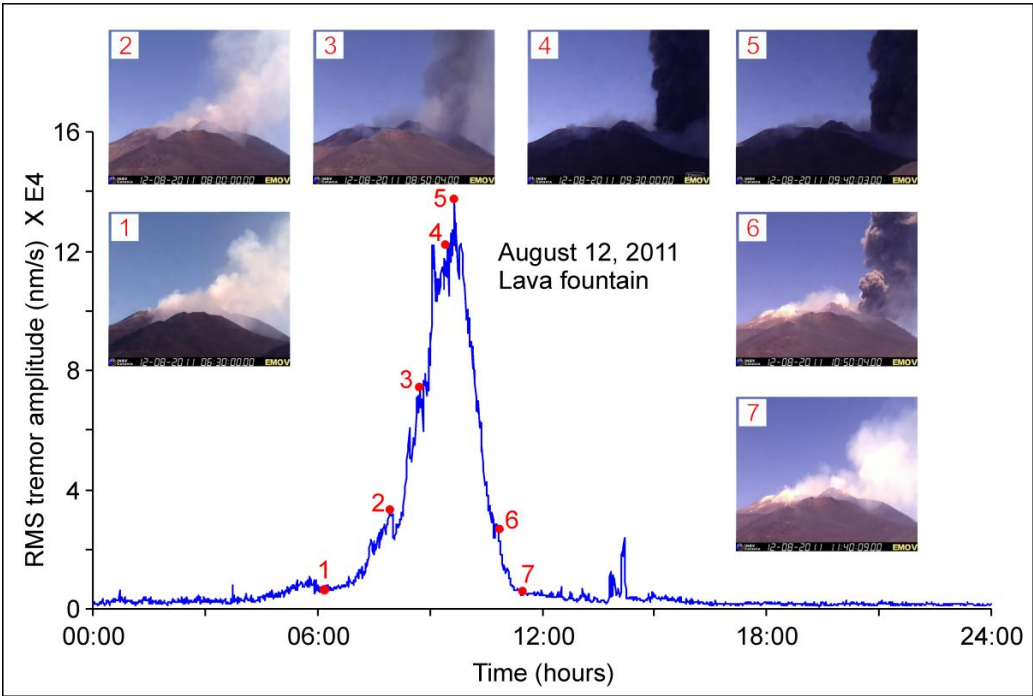


Figure 1. Example of RMS amplitude variations of volcanic tremor before, during and after a lava fountain. The paroxysmal episode shown here occurred on August 12, 2011. Numbered red dots match the sequence of snapshots taken from a video camera of INGV- OE. Given the fine weather conditions we are able to clearly relate the state of volcanic activity to the signal.

2. AMPLITUDE BASED MONITORING

2.1. Multi-Station Alert

More than 100 lava fountains occurred in 2000 and 2001. These permitted us to accumulate precious experience with respect to the development of volcanic tremor before, during and after paroxysmal events, allowing us to set up first rules for warning governmental authorities (see e.g., Alparone et al, 2003; Falsaperla et al, 2005). This experience turned out useful in particular in the time before the flank eruptions in July 2001 and October 2002, when the observations together with the analyses of earthquake swarms led to a rather clear picture with respect to the impending events. All these observations have encouraged the development of automatic alert systems for early warning purposes. The progress with respect to data quality and improving knowledge of signal characteristics now allows us to alert first INGV-OE personnel and in consequence authorities - like Civil Protection and the Tower of the Fontanarossa Airport - with considerable reliability at a very early stage of unrest.

One of the strategies we present here is based on the observation of changes of volcanic tremor amplitude (given in terms of ground motion velocity) with time. These changes depend on the status of volcanic activity as well as on the distance of receivers from seismic sources. The locations of the latter have been shown to vary (see, e.g., Di Grazia et al, 2006), which hinders the direct use of amplitude thresholds for warning purposes. On the other hand STA/LTA ratios offer an elegant way out of the problem (see Figure 2). Migration of the sources will affect both short term average STA and long term average LTA, unless they occur very rapidly – basically in times of the order or less of the LTA duration. Indeed, as experience has shown, the ratio between STA, calculated over one hour, and LTA, averaged over one day, remains more or less constant as long as no significant change of the volcano dynamics is encountered.

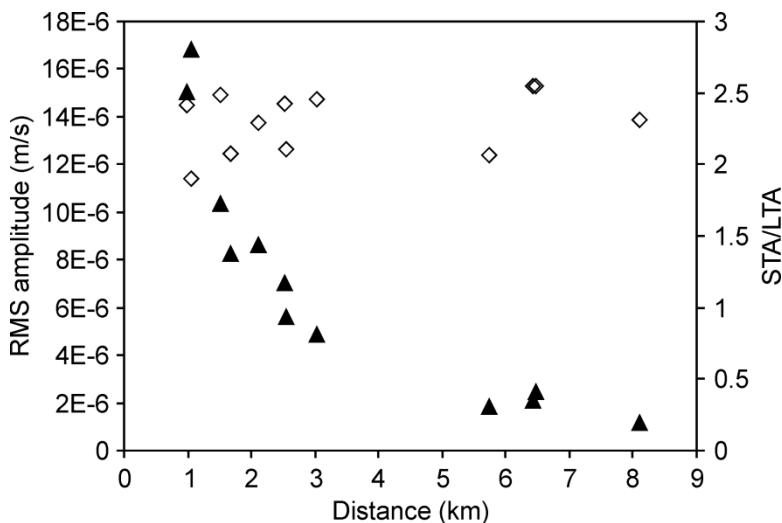


Figure 2. RMS amplitudes (triangles) and STA/LTA (diamonds) on April 10, 2011, 05:40 UT (four hours before a lava fountain). Notice the rapid decay of amplitudes with distance meanwhile the STA/LTA relation is rather constant. Most stations show STA/LTA > 2, i.e., a first alarm is issued.

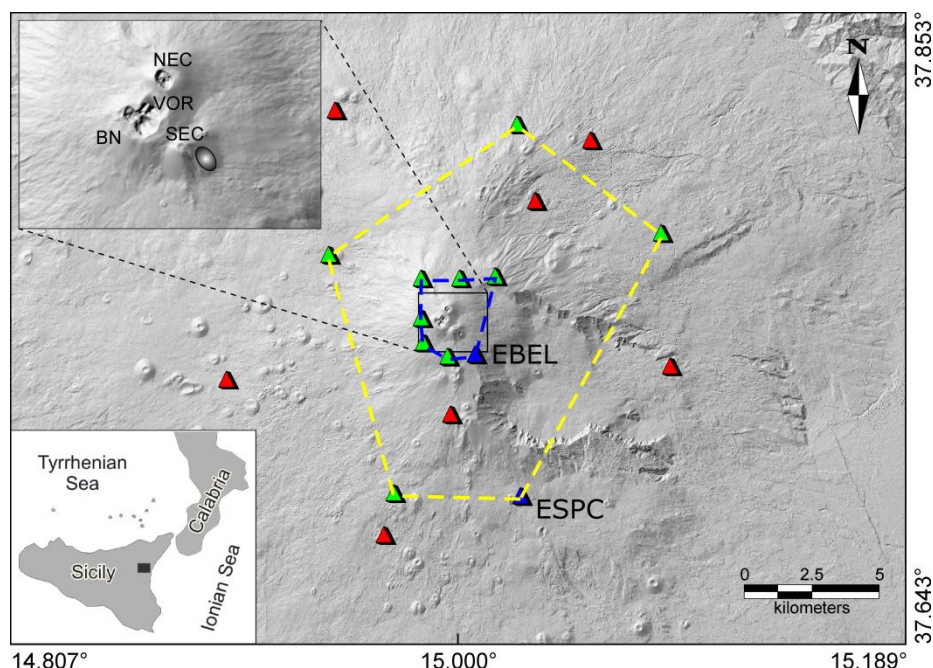


Figure 3. Positions of the nineteen key stations considered in this paper. The location of tremor sources is carried out using all stations. Criteria for RMS-based alert are obtained using stations marked as green and blue triangles (12), whereas on-line classification of tremor spectra are carried out for the two stations ESPC and EBEL (blue triangles). The inlet in the upper left corner shows the summit area with its four main craters (NEC=North-East Crater, BN=Bocca Nuova, VOR=Voragine, SEC=South East Crater). The peripheral ring (PR) is marked by dashed yellow line, the summit ring (SR) by dashed blue line.

It was therefore possible to base the design of a multi-station alert system upon the STA/LTA criterion taking advantage of the increasing density of the good quality seismic network running on Mt Etna. One of the advantages of this amplitude based multi-station approach resides in its stability, as it is not sensitive to the failure of single stations. At the same time, false alarms caused by noisy conditions at single stations can be widely avoided.

Currently, twelve key stations are considered (Figure 3). Their locations delineate two concentric rings: (i) a summit ring (hereafter SR) and (ii) a peripheral ring (hereafter PR), ranging between ca. 1 to 3 km and 5 to 8 km from Central Crater area, respectively.

In order to define the amplitude in a specific frequency band, we compute spectra applying 1024 points (10.24 s) Fast Fourier Transform (FFT), with overlapping of 24 points. From the power spectra we easily obtain the RMS amplitude in the frequency band of interest – here 1.5 to 5.5 Hz - by applying Parseval's theorem. We prefer this scheme to time domain calculus as the choice of frequency bands becomes rather flexible. Indeed, the raw time series have to be accessed only once for the FFT. Using a gliding window scheme with a 10s step width, ensembles of RMS amplitudes are formed, i.e., 360 values for the 1 hour STA window and 8640 for the 1 day LTA window. Thus, each value of STA and LTA – calculated every 10 s - is an average of 360 and 8640 RMS amplitudes, respectively. The choice of our frequency band follows the scope to limit the risk of false alarms. We exclude low frequencies which dominate spectra of teleseismic earthquakes as well as high frequencies which are sensitive to human activity or wind. Our experience has shown that in the above

mentioned frequency band, STA/LTA ratios greater than 2 often come along with a rapid increase of absolute amplitudes of volcanic tremor. During paroxysmal events the STA/LTA values reaches values even greater than 10.

In our alert scheme each station sets an alarm flag when the first trigger level ($STA/LTA > 2$) is exceeded 30 times consecutively, i.e., STA/LTA is above the threshold for five minutes. Having four stations with an $STA/LTA > 2$ a first alarm (level 1) is prompted to the personnel being present in the operative centre of INGV-OE. Following the protocols actually in use, a first alert is communicated to certain official authorities. Further, several tests brought out the need to introduce a second trigger level ($STA/LTA > 4$, maintained for five minutes). This level is definitely reached only during paroxysmal events, and becomes a crucial hint for violent volcanic activity even when this cannot be verified by means of field observation and/or from volcano imagery (for instance because of bad weather conditions, events during night time, failure of optical, thermal, and/or infrared cameras). The return to a pre-eruptive status is declared when the STA/LTA ratio value drops below 1. For each seismic station, all status changes must be confirmed uninterruptedly for at least three minutes.

Table 1. Paroxysmal episodes in 2011 and STA/LTA alert times

Date	Start time paroxysm	End time	Duration	Alert issue time	Lead Time
Jan 12	22:00	23:30	1:30	20110112_20:32	1:28
Feb 18	no visual observation		-	20110218_03:59	-
Apr 10	09:15/09:30	11:30	2:00/2:15	20110410_01:45	7:30/7:45
May 12	01:45	03:30	1:45	20110511_19:04	6:41
Jul 09	13:50	15:15	1:25	20110709_07:37	6:13
Jul 19	00:30	02:00	1:30	20110718_23:20	1:10
Jul 25	02:30	06:30	4:00	20110724_20:45	5:45
Jul 30	19:30	21:30/22:00	2:00/2:30	20110730_08:31	10:59
Aug 05	21:30	23:00	1:30	20110805_19:19	2:11
Aug 12	08:30	10:30	2:00	20110812_05:29	3:01
Aug 20	07:00	07:50	0:50	20110820_05:17	1:43
Aug 29	04:05	07:53	3:48	20110829_01:20	2:45
Sep 08	07:00	08:25	1:25	20110908_06:11	0:49
Sep 19	12:20	13:00	0:40	20110919_07:21	4:59
Sep 28	19:30	20:10	0:40	20110928_15:07	4:23
Oct 08	14:30	14:50	0:20	20111008_11:46	2:44
Oct 23	18:26	20:30	2:04	20111023_18:10	0:16
Nov 15	11:15	12:50	1:35	20111115_06:08	5:07

Besides these general rules, some additional precautions have been taken in order to limit false alerts. (i) among stations flagging a status change, at least one must belong to the SR and at least one to the PR; the SR stations alone are not sufficient to create an alarm. (ii) the average of RMS amplitude calculated for stations belonging to SR must be at least three times

the one of PR. These additional constraints aim at avoiding alerts for extremely local events or disturbances. Besides, false alerts due to seismic transients (such as swarms of shallow earthquakes, teleseism, etc.) are widely excluded.

The eighteen lava fountain episodes occurred on Mt Etna in 2011, provided rich material for further testing the automatic alert system in the context of volcano monitoring. The results of the alert system are summarized in table 1. On the whole, the multi-station alert system issued first level alarms several hours before the onset of paroxysm; on 30 July, this lead time was over 10 hours. An exception is represented by the October 23 lava fountain. In this case, volcanic tremor amplitude increased very rapidly, being indicative for a very fast magma uprise. Thus, the alert system produced the first level alarm only sixteen minutes before the start of the lava fountain. No comparison is possible for the February 18 eruptive episode as bad meteorological conditions prevented any volcanological visual observation at its beginning.

2.2. Volcanic Tremor Source Location

Changes of tremor characteristics are often accompanied by a migration of the volcanic tremor source. On the basis of the data recorded during the 2001 eruption, Falsaperla et al, (2005) found clear hints for considerable changes of depth and horizontal position of volcanic tremor sources during this eruption. This observation led to the development of a specific software for tremor source location, exploiting the amplitude distribution across the seismic network (Di Grazia et al., 2006). Since 2006 locations of the tremor source are carried out automatically and displayed in the INGV operative centre of Catania. The tremor locations have turned out to be extremely useful for warning purposes throughout both summit activity and flank eruptions, such as during the 2007 lava fountain episodes as well as the flank eruption that started on May 13, 2008 (Patanè et al., 2008; Di Grazia et al. 2009). In all these cases a clear migration of tremor sources towards the eruptive centres could be noticed hours before the activity became manifest at the surface. The location of the tremor source completes the picture of an imminent volcanic unrest, highlighting the magma pathway which feeds the eruptive activity. The dynamics inferred from tremor source migrations corroborate the early warnings flagged by the changes of the seismic signal characteristics.

Given the nature of the tremor signal, standard location techniques - based on the inversion of traveltimes of seismic phases - are not applicable. Therefore, we use a method taking into account the spatial distribution of RMS amplitudes. At the moment of writing the location is based on nineteen key stations, which belong to the broadband permanent seismic network run by INGV-OE. Such key stations are located at distance ranging between about 1 and 9 km from the centre of summit area (Figure 3).

Volcanic tremor source location is estimated assuming the propagation in a homogeneous medium and a seismic amplitude decay with distance (Gottschammer and Surono, 2000; Battaglia and Aki, 2003; Battaglia et al., 2005; Di Grazia et al., 2006). In a simplified description, seismic wave attenuation is controlled by two factors, i.e., geometrical spreading and absorption. The geometrical spreading is assumed to be inversely proportional to the source-receiver distance, in the case of body waves, or to the square root of such a distance (surface waves). The second factor, absorption, expresses the loss of energy as seismic waves

propagate. Usually the phenomenon is treated without attempting to separate the various mechanisms generating loss of energy, defining a quality factor Q (Knopoff, 1964), such that

$$\frac{1}{Q} = \frac{\Delta E}{2\pi E} \quad (1)$$

where ΔE is the elastic energy loss during each cycle and E is the peak energy reached in that cycle. By combining geometrical spreading and absorption, amplitude decay can be expressed according to the general law:

$$A(f, s) = A_0(f) s^{-b} e^{-\alpha s}, \alpha = \frac{\pi f}{Qc} \quad (2)$$

where f is the frequency, c the seismic wave velocity, b the exponent standing for the geometrical spreading and s is the distance source - receiver. For the reasons stated below we adopt $b=1$, i.e., a geometrical spreading corresponding to body waves. Q represents the ray-path-averaged quality factor. The equation (2) can be linearized taking the logarithm on both sides as:

$$\ln A_i + \alpha s_i = \ln A_0 - b \ln s_i \quad (3)$$

where A_i is the amplitude measured at the i_{th} station, A_0 is the reference amplitude of the signal generated at the source, α is the frequency-dependent absorption coefficient, and s_i is the corresponding source-to-receiver distance. In this form $\ln A_0$ and b represent the intercept and the slope of the linear equation, respectively.

RMS tremor amplitudes are obtained on the basis of the Fourier spectra using the scheme described in section 2.1, this time using the frequencies between 0.5 and 2.5 Hz, as these are the most significant ones of volcanic tremor radiation (Falsaperla et al., 2005). From every ensemble of 360 RMS amplitudes, covering 1 hour of tremor, the 25th percentile is extracted. This allows us to efficiently remove undesired transients in the signal and consider continuous recordings (Di Grazia et al., 2006).

In the original form of Di Grazia et al. (2006), the inversion was simultaneously carried out for both the source location and geometrical decay factor, using a systematic search approach where the source location was shifted over a 3D grid and the geometrical decay factor was obtained from linear regression of eq. (2). The authors adopted this approach in order to avoid any a-priori assumption about the wave type. As experience has shown, however, the estimated decay parameters (b in eq. 2) are close to 1, i.e., supporting body wave like decay for tremor amplitudes. This is in agreement with Saccorotti et al. (2004), who report a rather complex nature of the wave field of seismic tremor, however body wave characteristics prevailing in the frequency range between 0.8 and 2.3 Hz. Besides, test locations assuming a decay valid for surface waves ($b=0.5$) led always to source positions at the very surface, which is a strong hint for an artefact. The goodness of fit was significantly smaller for surface wave decay than for body wave decay. As a consequence the method applied here uses – similar to Di Grazia et al. (2009) – a constant geometrical decay parameter

of $b=1$ (body waves), which reduces the degree of freedom of the inversion and avoids the risk of fitting noise.

Following Battaglia and Aki, (2003), the frequency-dependent absorption factor should not have a large influence on the source locations. As a precaution we include the absorption parameter α of eq. (2) in our grid search approach, varying it with a step width of 0.01 in a range of $0 \leq \alpha \leq 0.4$. The best values of goodness of fit (R^2 hereafter) are achieved with very low α values, confirming the findings of Battaglia and Aki (2003).

RMS amplitudes are calculated as vector sums of the three components recorded by 3C seismometers, rather than using only one component as in Di Grazia et al. (2006). This further improves the stability of locations.

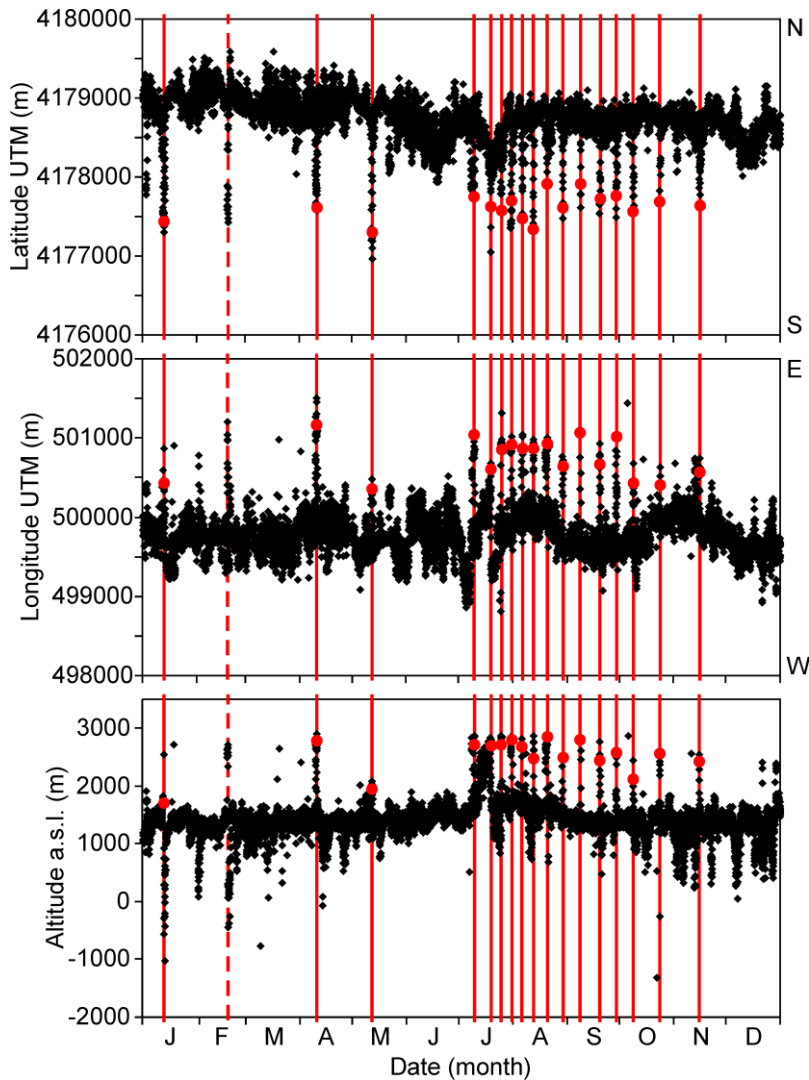


Figure 4. Tremor source locations during 2011. Occurrence of paroxysmal episodes (lava fountains) is marked with a red line, dashed for 18 February, which was not visible at its beginning because of bad weather conditions. The red dots mark the position of the tremor source at the beginning of activity visible at the surface.

The source location of tremor is established in a systematic search, moving the source over a 3D grid and calculating the R^2 encountered for each grid point. The grid is centred underneath the summit craters (Figure 3) and has a dimension of 6 x 6 x 6 km in the two horizontal and the vertical directions. The spacing of the grid points is 250 m. Through the 3D grid search, we identify the most probable source locations from points where the measured amplitudes best fits the amplitude decay law in eq. (2), i.e., R^2 are maximal. The final source location is obtained as the centroid position of all 3D grid points whose R^2 do not differ more than 1% with respect to the maximum of R^2 (Di Grazia et al., 2006). We reject low quality solutions accepting a result only when: i) the goodness of the R^2 fit is ≥ 0.9 , and ii) at least 13 out of the 19 stations are available.

Figure 4 gives an overview of tremor source locations in 2011. Most of the time the sources are located right underneath the centre of the summit area of Mt Etna, and the vertical position is found typically at 1000-1500 m a.s.l., i.e., ca. 1500-2000 m below the summit area. Before and during a lava fountain event (indicated by red lines in the figure) we notice a rapid movement in SE direction, towards the SE and New SE crater (SEC). At the same time, we also notice that sources reach practically the surface. This is consistent with visual observations and infrasound records, and confirms the reliability of the location method.

Besides short lived episodes of source migrations in the context of paroxysmal events, we notice long lasting trends, in particular during summer 2011. Towards the end of June / beginning of July sources are found at more and more shallow levels, even through periods of relative quiescence. After the sources reached the surface during the lava fountains on July 9 and 19, 2011, paroxysms become rather frequent, with time intervals of one week or less. In the second half of the year, sources return to greater depth, similar to those noted at the beginning of 2011, and time intervals between paroxysms increase.

3. PATTERN RECOGNITION USING UNSUPERVISED CLASSIFICATION

Continuous acquisition of seismic waveforms generally comes along with the accumulation of huge amounts of data every day. Data reduction and parameter extraction are basic steps in order to facilitate the handling and interpretation of large data volumes. As a consequence single valued data streams, such as seismograms, are transformed to objects, described by multi-dimensional feature vectors (see Appendix A1). The suitable representation of the object's characteristics is a critical issue, and their development with time is of particular interest for our purposes. A possible way to cope with this problem resides in automatic classification of the signals (see e.g., Falsaperla et al., 1996; Langer and Falsaperla, 1996; Esposito et al., 2008). In automatic classification techniques a data set is typically given by a table with rows and columns. By convention, a row corresponds to a single pattern or feature vector and columns represent its components. As continuous time series of seismic data are not suitable to be used as feature vectors, spectral representations are preferred. Langer et al. (2009) analyzed spectral patterns derived from volcanic tremor data recorded during the July-August 2001 flank eruption of Mt Etna, applying both concepts of supervised (Support Vector Machines and Multilayer Perceptrons) and unsupervised classification (Self Organizing Maps and crisp clustering). They were able to identify various

regimes of patterns matching volcanological observations. The results of the classification with supervision confirmed the a-priori distinction of different regimes of volcanic activity, i.e., pre-eruptive, eruptive, post eruptive and lava fountains (Masotti et al., 2006). On the other hand, unsupervised classification brought out subtle internal structures within the data, adding valuable information to the results of the a-priori classification. Encouraged by the results of Langer et al. (2009), Messina and Langer (2011) developed a software package for unsupervised classification (freely available at the URL <http://earthref.org/ERDA/974/>), which uses the so-called Kohonen maps as well as various techniques for clustering.

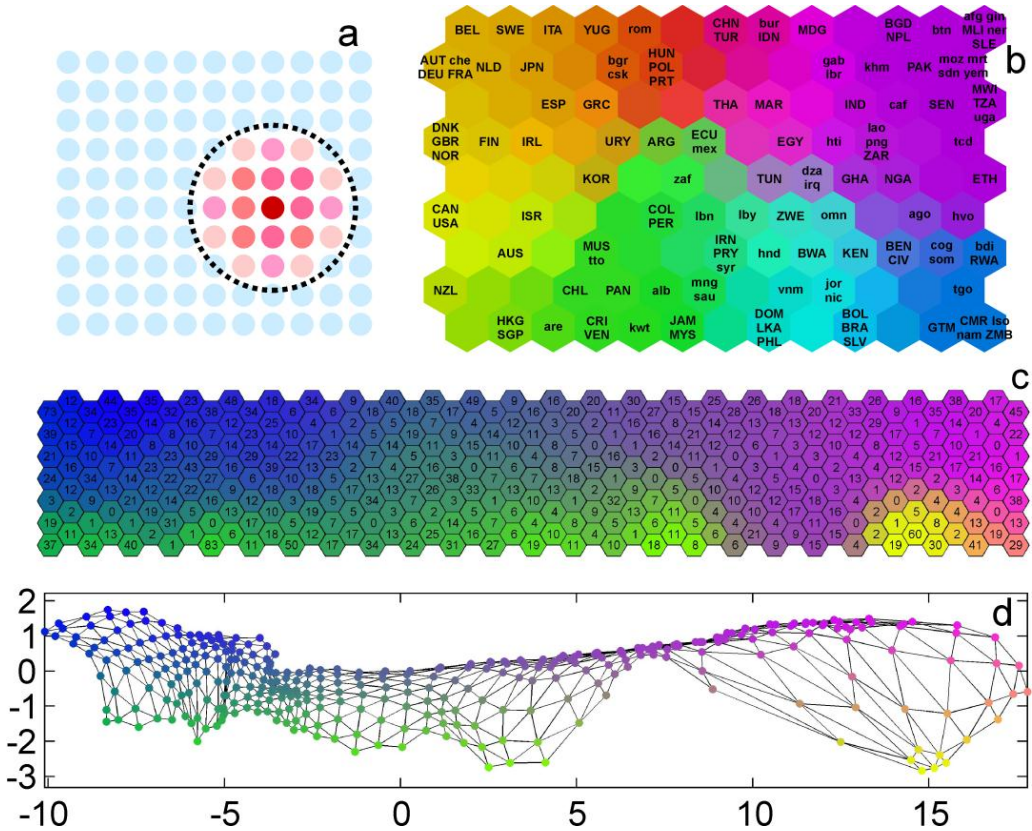


Figure 5. The Self Organizing Map (SOM). (a) During the training process, values of the nodes of the SOM are compared to the feature vectors and adjusted with a specific learning rate to minimize the difference. Not only the node being closest to a pattern is updated, but also the neighboring ones are adjusted although with a lower learning rate. (b) SOM of the World Poverty Map created on the basis of 39 socio-economic parameters measured worldwide. From the colors assigned to the countries and their position on the map, we clearly distinguish lesser developed countries and more advanced ones. Though belonging to the advanced countries, clear differences between countries in Continental Europe and North America are evident. (c) 2D map of our 5500 62-dimensional spectral patterns, representing neighborhood relations schematically. Small numbers on the hexagons report the number of patterns for which the nodes were identified as “Best Matching Unit” (BMU). (d) 2D projection of weights of the SOM in a system of axes spanned by the two major Principal Components. Note that Principal Components are formally dimensionless. Whereas the neighborhood relations among the SOM nodes are schematically depicted in Figure 4b, c, distances between nodes are represented more faithfully in (d).

One of the clues of the approach is the synopsis of various unsupervised classification techniques. Similarly to Langer et al. (2011), we focus here on Kohonen maps (Kohonen, 1984, 2001) and fuzzy clustering (Zadeh, 1965). Kohonen maps, or “Self Organizing Maps” (SOM hereafter) are made up of a number of nodes, each one represented by a feature vector and being representative of a number of patterns. A pattern in our application is given by the power spectral densities in the frequency bins. During an iterative procedure - the so-called training phase - the node weights (feature vector elements) are adjusted minimizing the differences between the nodes and the original pattern (Appendix A1.1, eq. A1 and A2). A pattern is assigned to the node for which the smallest difference is encountered; this node is referred to as the “Best Matching Unit” (BMU) for that pattern (Figure 5a). After the training phase, the nodes are mapped to a 2D representation space - the “map” - using Principal Component Analysis. Figure 5b is an example proposed by Kohonen (see Kohonen, 2001) grouping some 190 countries with respect to their socio-economic parameters. Countries with similar characteristics are found close to each other on the map and belong to BMU’s with similar colors. In Figure 5c we apply Kohonen’s idea to our spectral patterns. We schematically represent the position of the nodes in a mesh of hexagons. How many of our patterns belong to a BMU can be read from the number within each hexagon. Nodes marked with a ‘0’ are “looser nodes”, which are not recognized as BMUs by any pattern. An appealing property of SOM is the topological fidelity, i.e., patterns represented by neighboring nodes in the SOM are close to each other also in the original data space. Analog to the example shown in Figure 5b we apply a color coding to the nodes depending on their position on the map and assign to each pattern the color of the BMU to which it belongs. Thus, given the topological fidelity similar patterns show similar colors.

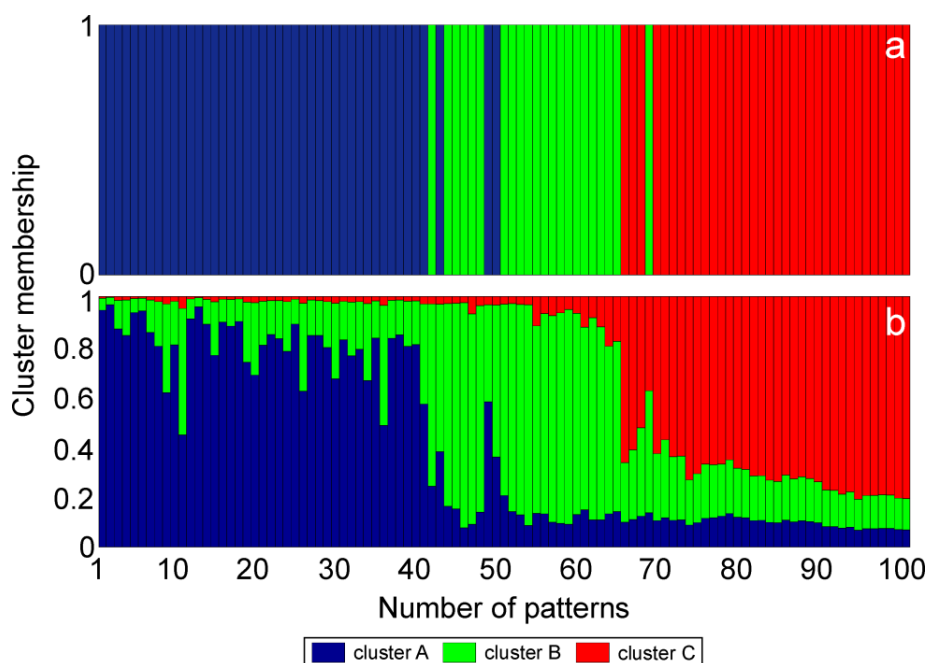


Figure 6. Crisp and fuzzy clustering. In crisp clustering (a) each pattern is assigned exclusively to a class, whereas in fuzzy clustering (b) the class membership is a vector whose components express to which degree a pattern belongs to a certain class (from Langer et al., 2011, modified).

Along with SOM creation, a fuzzy clustering algorithm is exploited in our approach. In Fuzzy Clustering (Figure 6) the class membership of a pattern is given by a vector rather than by a single ID. Vector components indicate to which degree a pattern belongs to the various clusters, in order that the sum of these membership rates is 1. A pattern with a cluster membership 1 belongs exclusively to one cluster, equivalent to a crisp membership. A prevailing membership of 0.9 is close to crisp, whereas < 0.5 means rather fuzzy.

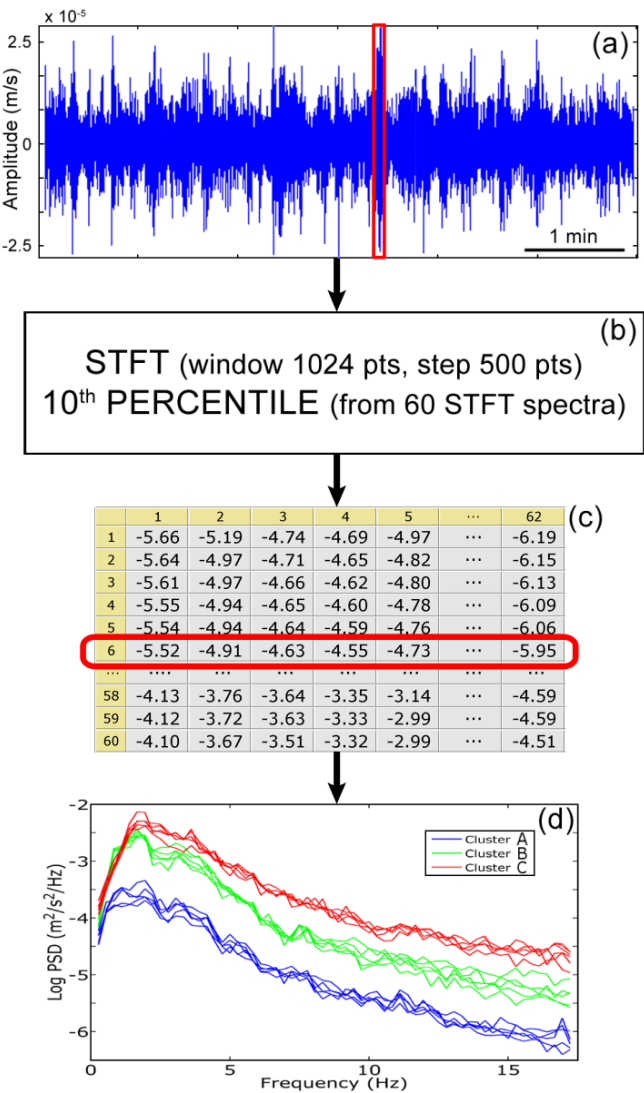


Figure 7. Preprocessing steps for the creation of patterns. (a) Typical record of volcanic tremor obtained at station ESPC; the red box marks a time window of 1024 points. (b) Short Time Fourier Transform (STFT) carried out in a gliding window scheme, moving the window in time steps of 5 s. (c) Sorted ensemble of 60 individual spectra obtained by STFT. Each column in the table corresponds to a frequency bin with a width of ~0.29 Hz. The 6th row, marked in red, corresponds to the 10th percentile spectrum, which forms the feature vector describing the pattern. (d) Examples of patterns used in our classification. Colors are related to the prevailing cluster membership encountered in fuzzy cluster analysis.

Visualization is a critical issue in multivariate data analysis. The colors of the nodes of the SOM and the class membership vectors inferred from fuzzy clustering offer an elegant way for monitoring the development of volcanic tremor characteristics. This kind of synoptic view proved indeed to be especially useful, given the large number of patterns, allowing the identification of transitional regimes between clusters, which are of paramount interest for surveillance purposes.

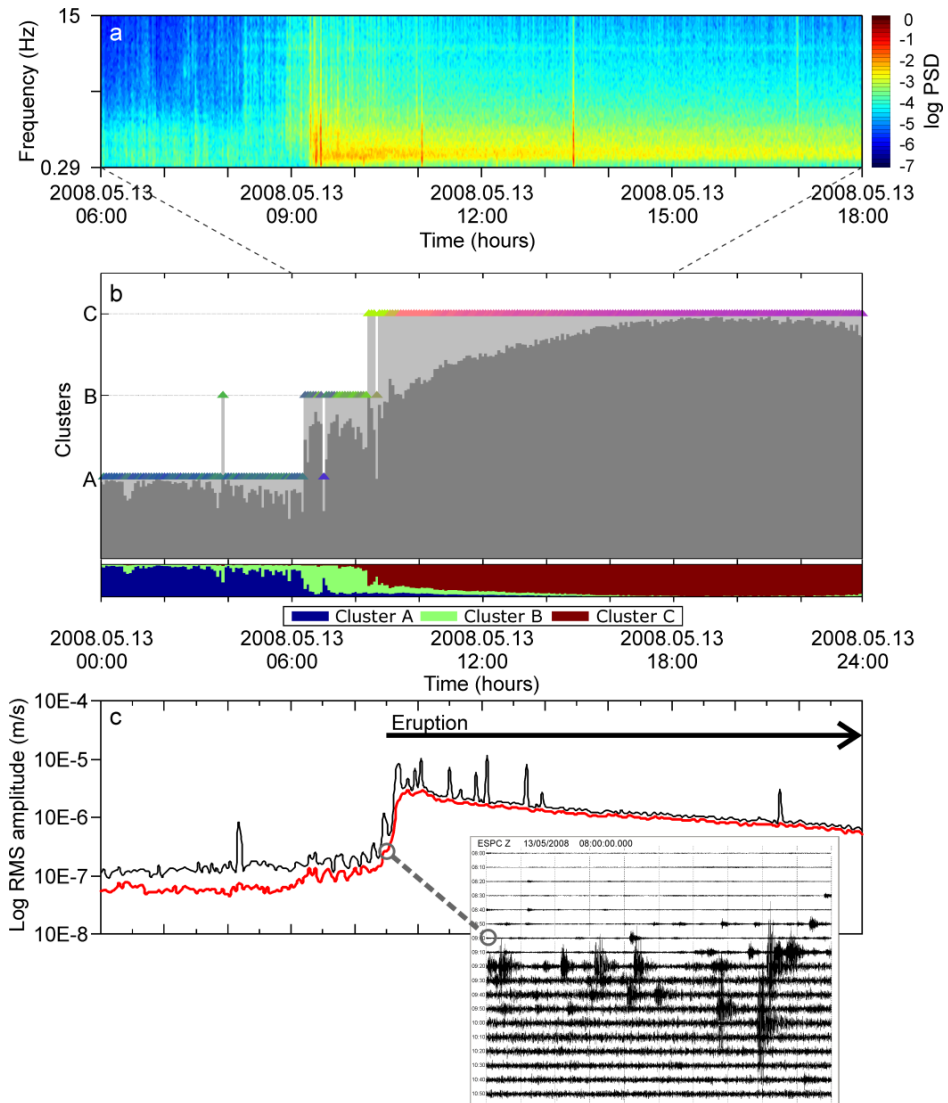


Figure 8. Spectrogram (a), synoptic classification of tremor spectra (b), and time series (c) for the day of 13 May 2008. The dark curve in (c) corresponds to the averages calculated over 30 s, whilst the red one represents 10th percentiles in windows of 5 min. Note that the use of the percentiles removes the influence of transient signals, such as earthquakes, which are frequent at the beginning of volcanic unrests (see seismogram with the onset of the seismic swarm in the inset). The dark arrow indicates the time of manifest eruptive activity during the day of May 13, 2008 (from Langer et al., 2011, modified).

We define the spectral patterns pre-processing the data in a similar way as described in section 2.1, where spectrograms are calculated applying a gliding window scheme (see Figure 7). In this scheme a Short Time Fourier Transform (STFT) is carried out over 1024 points. The time window is shifted of 500 points (i.e., 5 seconds). For each 5 s time step we obtain a preliminary feature vector made up by spectral amplitudes measured in equally spaced frequency intervals. We then consider ensembles of 60 preliminary feature vectors. Again we use low percentiles (10th percentile here) instead of averages, getting – at least partially – rid of transients (e.g., wind gusts, earthquakes, etc), which are considered as disturbance in tremor data analysis (Di Grazia et al., 2006). Finally, we have one pattern every five minutes. To limit the dimension of the feature vector, frequency bins of ca. 0.29 Hz - representing the spectral content in successive frequency intervals - are formed. At the end of the process we obtain feature vectors with 62 spectral components, covering a frequency range up to ca. 18 Hz.

Stability of pre-processing and affordable computational effort have led us to set up a real-time system for the simultaneous application of the SOM and fuzzy clustering for monitoring and early warning purposes. We followed a concept outlined in Langer et al. (2011), who considered data recorded in the context of eight eruptive events (seven lava fountains and one flank eruption) in 2007 and 2008. They demonstrated that subtle but significant changes in the tremor characteristics occur well before the onset of manifest activity (Figure 8). The authors proposed to use this data set as reference, as it encompasses a large variety of eruptive stages such as quiescence, lava fountains, flank eruptions as well as transitional stages before and after an unrest.

In the online application, incoming new data are stored in a ring buffer, which covers 24 hours of the most recent signals of a single station and is updated every 5 minutes. Data in the ring buffer are pooled together with the reference data set, allowing us to compare the characteristics of the current signal to those recorded under known scenarios of volcanic activity. The experience collected during a test phase, covering more than two years, allows a better understanding how the system behaves under various conditions, both regarding the state of volcanic activity as well as environmental conditions (man made noise, unfavorable meteorology, such as wind, and probable changes in magma dynamics – e. g., intrusions in shallow levels – not leading to manifest activity at the surface).

In Figure 9 we show some examples of tremor classification for the station ESPC. They pertain to the absent or low volcanic activity. As the classification for the reference pool remains stable, it is shown only once (Figure 9a).

The panel shown in Figure 9b refers to a day with very quiet conditions (i.e., absent volcanic activity). Only patterns belonging to cluster 'A' are present, and SOM colors are dominated by blue and dark green tones. In the record shown in Figure 9c we notice again a prevailing membership to cluster 'A' and SOM colors with blue and green tones. During daytime, however, the cluster membership flips between 'A' and 'B'. This flipping trait is frequently met during daytime but is absent at night, and anthropogenic noise is likely to be the source of the phenomenon.

In Figure 9d we observe regimes lasting several hours, in which there are patterns belonging to cluster 'B' or even 'C' for a short time. SOM colors are green. Flipping cluster memberships (compare Figure 9c), are limited or absent. The phenomena occur both during day and night, so we exclude anthropogenic noise as their source. On the other hand, checking for meteorological conditions, we found that these patterns are typical of days with

strong wind with W or SW direction. In the example of Figure 9d, wind speed reached up to 60 knots at 3000 m a.s.l., an altitude slightly less than that of Mt Etna's summit. Figure 9e shows the results of the classification for the patterns recorded on April 18, 2007. In this case, we note a change starting at 03:00 UT, leading to a stable regime with cluster 'B', whereas SOM colors get violet and purple tones. After checking the weather conditions, we can exclude strong wind as a possible source.

The eighteen episodes of eruptive activity observed during 2011, represent a formidable test set for checking the validity of the classifier and for setting up criteria which allow us to issue automatically an early warning with a reasonable degree of reliability.

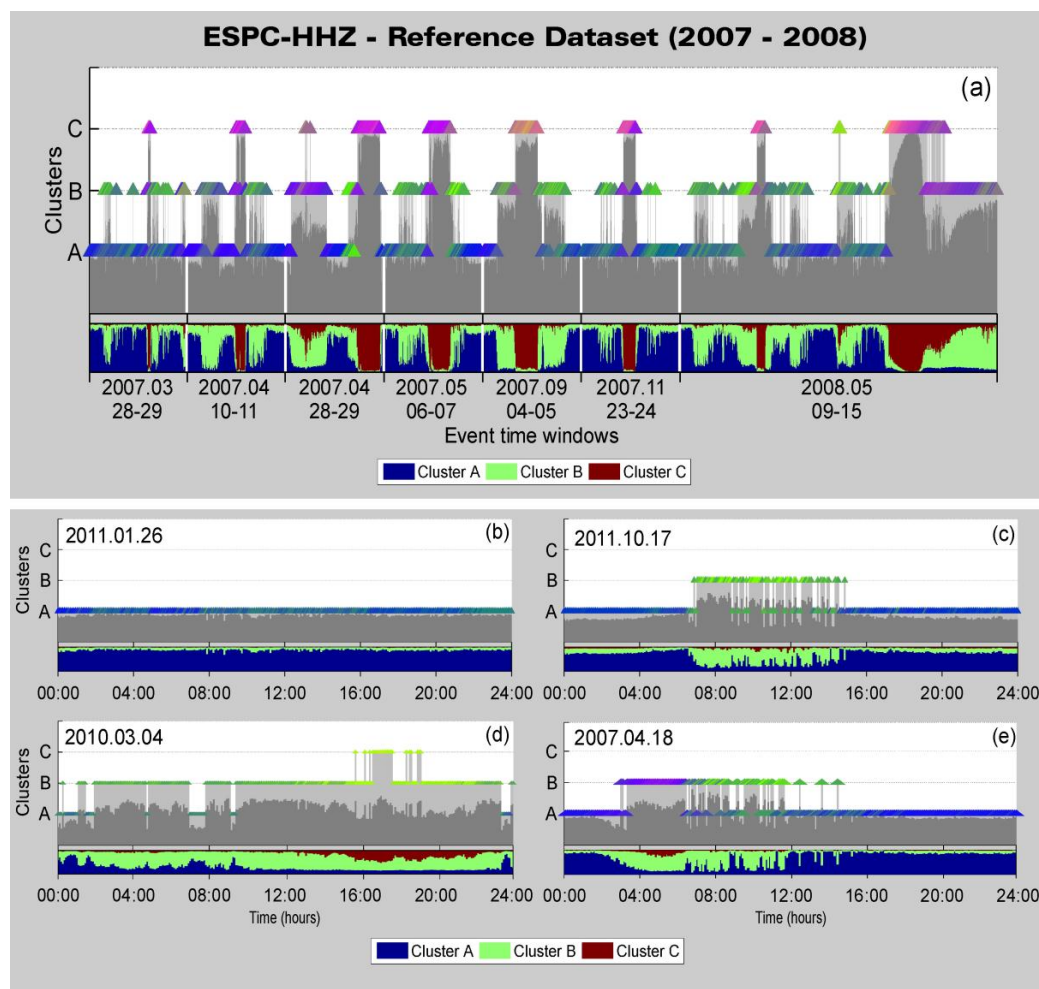


Figure 9. Examples of unsupervised classification pertaining to the absent or low volcanic activity. The station considered here is ESPC. SOM colors and cluster membership vectors of the reference data set are shown in (a). b–d are examples corresponding to: a day with very quiet conditions (b); anthropogenic noise during daytime (c); a day with very strong wind (d). The case shown in (e) is representative of the pattern classification of ca. twenty episodes of presumably failed paroxysms (possibly intrusions) observed in the time span between 29 March and 29 April, 2007.

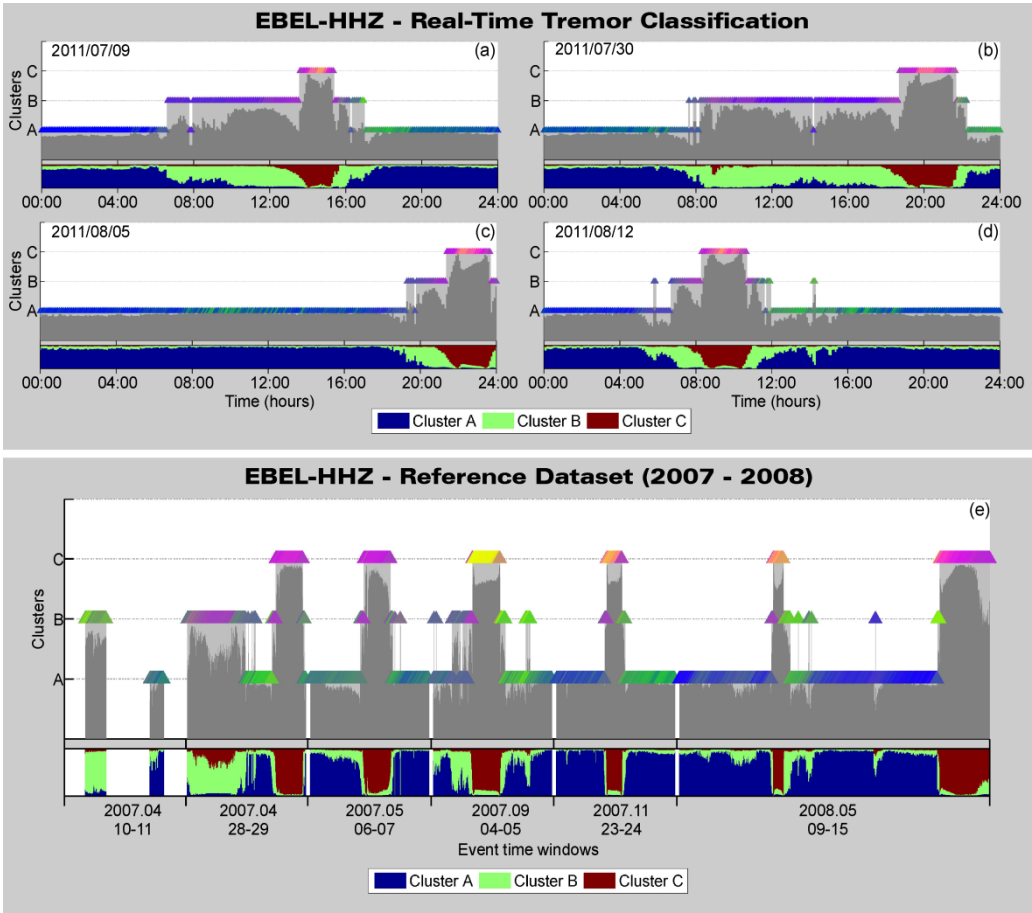


Figure 10. (a-d) Classification results for four episodes encountered during 2011 at the station EBEL, and (e) the reference data set, covering seven episodes in 2007 and 2008. The reference data set of EBEL is slightly poorer than the one for ESPC, on the other hand EBEL is less affected by noise.

Inspecting the results of the classifier (see Figure 10a-d) we notice a change of SOM colors in the time before the onset of an eruption. In particular, blue tones, which are typical for periods of quiescence, change towards lilac and violet. Approaching the climactic phase, the contribution of red tones becomes evident. The change of signal characteristics during the development of an unrest – here visualized by the SOM colors which become increasingly red - turned out to be stable over all of the eighteen 2011 unrests encountered so far. On the other hand, we also noticed that the green tones are present either during noisy conditions (such as strong wind, teleseisms) or in the coda of an unrest. In order to formalize a set of rules for early warning purposes, we therefore decided to analyze the development of red and green tones (Figure 11).

The diagrams shown in Figure 11 express the colors of the BMUs to which a pattern belongs, in numerical terms, and can be used for the definition of thresholds and additional criteria, such as how many times a condition must be “true” in a certain time window, before an alarm is flagged. The sensitivity of the settings depends on the rate of false alarms we are willing to admit. With a sensitive setting, such as reported in Tab 2a for station EBEL, an automatic alert will be issued well before the paroxysm becomes manifest at the surface (see

Tab 2b). For the unrest on Aug 29, 2011 our system behaves as shown in Figure 12. According to Tab 2b, criticality started at 00:35 UT and was confirmed for three consecutive patterns, which cover 15 min. Consequently, using the settings reported in Tab 2a an alarm is automatically issued at 00:35, a time which is indicated approximately by the red arrow in Figure 12. We notice that criticality is detected indeed at a very early stage of the unrest, where little or practically no change can be noticed on the original seismogram.

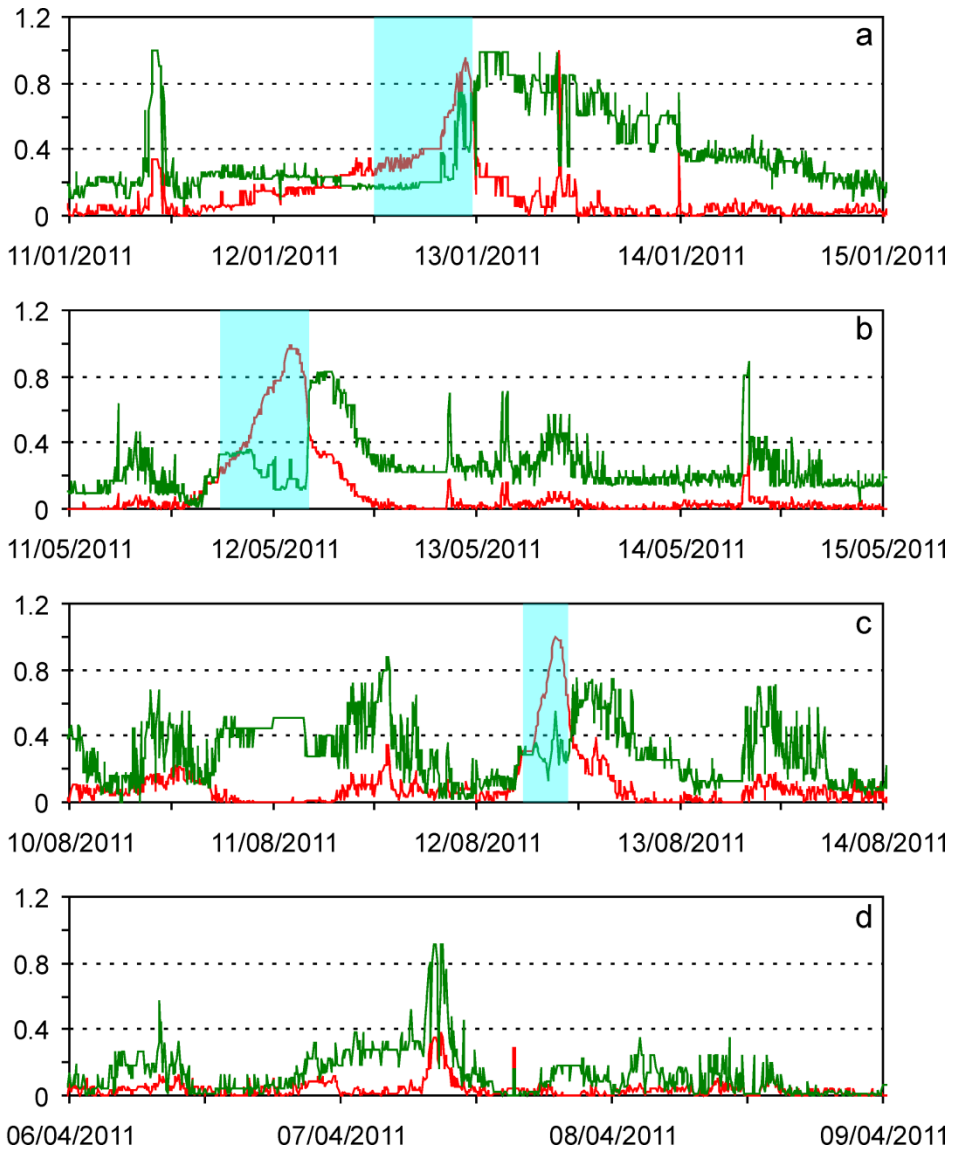


Figure 11. Development of red-green saturation for selected time intervals in 2011. Eruptive events are marked by the light-cyan rectangles. The lowest panel is related to a time span where no eruptive activity occurred. Contrary to eruptive episodes shown in panels a-c saturation of green in d) is larger than that of red.

Table 2a. Parameters for automatic alarm of the classifier (sensitive settings)

Parameter	Value	Meaning
FlagsNum	3	3 consecutive flags are considered
DiscToll	0	0% of FlagsNum flags may be 'false'
MinRedLevelActiv	0.2	Minimum level of red saturation for activation alert
MaxRedLevelActiv	1.0	Maximum level of red saturation for activation alert
RedGreenOffsetActiv	-0.25	Maximum difference of red-green saturation for alert
RedGreenOffsetDisactiv	-0.1	Maximum difference of red-green saturation for alert closing
MinRedLevelDisactiv	0.0	Minimum level red saturation for alert closing
MaxRedLevelDisactiv	0.30	Maximum level red saturation for alert closing
MinClusterActiv	A	Cluster required for alert
MaxClusterDisactiv	B	Cluster allowed for closing an alert

Table 2b. Automatic alarms by the tremor classifier (sensitive settings)

StartTime	EndTime	No patterns	Observations
20110102_20:45	20110103_03:50	85	Strong strombolian activity
20110112_08:05	20110113_00:25	196	Lava fountain
20110113_09:40	20110113_10:05	5	Lava fountain
20110113_10:50	20110113_11:05	3	Lava fountain
20110218_03:45	20110218_14:15	126	Lava fountain
20110218_14:30	20110218_14:50	4	Lava fountain
20110309_03:30	20110309_08:00	54	Teleseism
20110311_06:00	20110311_22:35	199	Teleseism (Japan M=9)
20110409_14:20	20110410_15:00	296	Lava fountain
20110511_17:35	20110512_07:10	163	Lava fountain
20110522_04:35	20110522_04:50	3	Light unrest, enhanced SO2
20110522_05:05	20110522_05:20	3	Light unrest, enhanced SO2
20110522_05:55	20110522_06:10	3	Light unrest, enhanced SO2
20110522_06:25	20110522_06:40	3	Light unrest, enhanced SO2
20110522_07:00	20110522_07:15	3	Light unrest, enhanced SO2
20110522_07:30	20110522_07:45	3	Light unrest, enhanced SO2
20110522_08:10	20110522_08:25	3	Light unrest, enhanced SO2
20110522_08:40	20110522_08:55	3	Light unrest, enhanced SO2
20110522_09:10	20110522_09:25	3	Light unrest, enhanced SO2
20110522_09:40	20110522_09:55	3	Light unrest, enhanced SO2
20110522_10:10	20110522_10:25	3	Light unrest, enhanced SO2
20110522_10:40	20110522_10:55	3	Light unrest, enhanced SO2
20110522_11:30	20110522_11:45	3	Light unrest, enhanced SO2
20110522_12:25	20110522_12:40	3	Light unrest, enhanced SO2
20110522_12:55	20110522_13:10	3	Light unrest, enhanced SO2
20110522_13:35	20110522_13:50	3	Light unrest, enhanced SO2
20110522_14:05	20110522_14:20	3	Light unrest, enhanced SO2
20110522_14:50	20110522_15:05	3	Light unrest, enhanced SO2

StartTime	EndTime	No patterns	Observations
20110522_15:40	20110522_15:55	3	Light unrest, enhanced SO ₂
20110706_20:30	20110708_07:35	421	Strong strombolian activity
20110709_06:10	20110709_17:10	132	Lava fountain
20110718_20:40	20110718_21:00	4	Lava fountain
20110718_21:20	20110719_03:05	69	Lava fountain
20110724_19:55	20110725_07:20	137	Lava fountain
20110728_19:45	20110728_20:00	3	?
20110730_06:35	20110730_06:50	3	Lava fountain (see text)
20110730_07:05	20110730_07:20	3	Lava fountain (see text)
20110730_07:35	20110730_22:10	175	Lava fountain (see text)
20110805_18:30	20110806_00:50	76	Lava fountain
20110806_01:05	20110806_01:20	3	Lava fountain
20110807_18:10	20110807_18:50	8	?
20110810_12:55	20110810_16:45	46	?
20110812_05:10	20110812_12:05	83	Lava fountain
20110812_15:25	20110812_15:40	3	Lava fountain
20110820_02:10	20110820_10:55	105	Lava fountain
20110829_00:35	20110829_05:45	62	Lava fountain (see text)
20110908_05:45	20110908_10:05	52	Lava fountain
20110919_08:00	20110919_14:00	72	Lava fountain
20110928_15:45	20110928_20:45	60	Lava fountain
20110928_21:00	20110928_21:15	3	Lava fountain
20110928_21:30	N.A		End of data encountered
20111023_10:50	20111023_11:20	6	seismic sequence
20111115_08:30	20111115_13:15	57	Lava fountain
20111203_08:10	20111203_09:30	16	?

Note: On Oct 23, 2011 program was down in afternoon hours.

Considering the year 2011, and settings reported in Tab 2a, we get ca. 35 alarm flags. Most of them are noticed in the context of a lava fountain unrest. In certain cases, for instance during the 30 July, 2011 lava fountain, the first alarms are repeatedly closed and re-opened, before a stable and lasting criticality is signaled by the system. We learn from this that the onset of the unrest was intermittent at its very beginning. We also notice false alarms which are marked as “?” in the table. The alarms signaled during May 22, 2011 are somewhat obscure: no vivid volcanic activity is reported in the days around May 22, however an enhanced emission of SO₂ is reported. On September 28, the system recognized correctly a criticality – there was a lava fountain, which culminated on 19:50 – but the system failed to close the alarm as no data were available.

It’s a matter of taste whether to accept the uncertainties of an automatic alarm system as reported in Tab 2b. A more robust behavior is found with the settings outlined in Tab 2c. We require a somewhat higher degree of red tones (0.25 instead of 0.20) and we request that 5 out of 7 consecutive patterns show criticality. In Tab 2d we see alarms only in the occasion of lava fountains with only one exception: the March 11, 2011 Japan earthquake which had a magnitude of 9. On the other hand we lose lead time: for instance, on July 30, the first pattern exhibiting criticality is encountered at 7:35 (instead of 6.35 with the more sensitive settings in

Tab 2a). As we have to consider 7 patterns, the alarm will be flagged only at 8:10. On August 29, the first critical pattern occurs at 1:05, and the alarm would be flagged at 1:40, more or less when in Figure 12a prevailing cluster membership flips from ‘A’ to ‘B’.

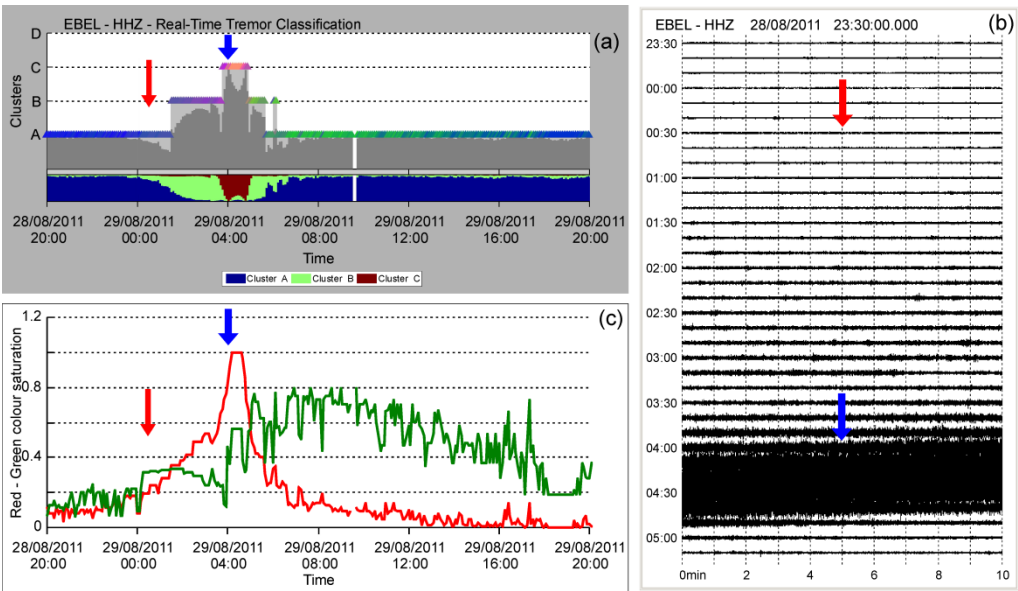


Figure 12. Behavior of the classifier during the unrest on August 29, 2011. The upper left panel shows the image displayed in the monitor room at INGV-OE; each colored triangle corresponds to a pattern representing 5 min. of tremor. The results of fuzzy clustering are also shown. The panel below shows the degree of red-green saturation, whereas the panel on the right hand side depicts the original time series. The red arrows indicate the time when automatic alarm was issued, the blue ones the onset of the lava fountain.

Table 2c. Parameters for automatic alarm of the classifier (robust settings)

Parameter	Value	Meaning
FlagsNum	7	7 consecutive flags are considered
DiscToll	20%	20% of FlagsNum flags may be ‘false’
MinRedLevelActiv	0.25	Minimum level of red saturation for activation alert
MaxRedLevelActiv	1.0	Maximum level of red saturation for activation alert
RedGreenOffsetActiv	-0.25	Maximum difference of red-green saturation for alert
RedGreenOffsetDisactiv	-0.1	Maximum difference of red-green saturation for alert discativation
MinRedLevelDisactiv	0.0	Minimum level red saturation for alert disactivation
MaxRedLevelDisactiv	0.30	Maximum level red saturation for alert disactivation
MinClusterActiv	A	Cluster required for alert
MaxClusterDisactiv	B	Cluster allowed for closing an alert

Table 2d. Automatic alarms by the tremor classifier (robust settings)

StartTime	EndTime	No Patterns	Observations
20110102_21:10	20110103_03:50	80	Strong Strombolian activity
20110112_09:30	20110113_00:25	179	Lava fountain
20110218_03:45	20110218_14:50	133	Lava fountain
20110311_06:00	20110311_22:35	199	Teleseism Japan M=9
20110409_20:25	20110410_15:00	223	Lava fountain
20110511_18:25	20110512_07:10	153	Lava fountain
20110708_00:10	20110708_07:20	86	Strong Strombolian activity
20110709_06:10	20110709_17:10	132	Lava fountain
20110718_22:00	20110719_03:05	61	Lava fountain
20110724_19:55	20110725_07:20	137	Lava fountain
20110730_07:30	20110730_22:10	176	Lava fountain
20110805_18:45	20110806_00:50	73	Lava fountain
20110812_05:20	20110812_12:05	81	Lava fountain
20110820_02:10	20110820_10:55	105	Lava fountain
20110829_01:05	20110829_05:45	56	Lava fountain
20110908_05:45	20110908_10:05	52	Lava fountain
20110919_08:00	20110919_14:00	72	Lava fountain
20110928_15:45	20110928_20:45	60	Lava fountain
20111115_08:30	20111115_13:15	57	Lava fountain

4. TECHNOLOGICAL SOLUTIONS

4.1. Volcanic Tremor in the Context of Multi Parameter Volcano Monitoring

On-line data processing, eventually leading to the creation of automatic alerts, critically depends on computational efficiency, robustness of the methods as well as the reliability of the data acquisition system.

Given the large number of stations deployed on Mt Etna and surrounding areas (110 seismic stations, 72 GPS, 23 tiltmeters, 17 gas measurement devices, 9 multiparametric and 10 cameras for imagery in the visible and infrared range) the amount of continuously recorded data on Mt Etna is huge, such as 20 GB per day. Besides, the type of data is quite different depending on the discipline to which they are related. Most of them represent series of real numbers (e.g., seismograms and derived parameters, infrasound, deformation, temperature, gas measurements).

On the other hand, data obtained from camera monitoring are given as digital images in form of single snapshots or videos. Even though we are dealing here essentially with tremor, i.e., seismic data, information from other disciplines is extremely important as it helps to set up valid criteria for early warning purposes and automatic alert issues.

In a strict sense, seismic data yield only indirect evidences of a volcanological phenomenon, which needs to be confirmed by other observations providing the proof of “the

smoking gun”¹. In particular we are interested to understand whether changes in the characteristics of tremor are related to volcano dynamics or simply are caused by a disturbance, such as human activity, meteorological effects or earthquakes (which are undesired signals in our context).

4.2. Data Acquisition Issues

The use of innovative techniques for the processing of real time data requires high performance systems in terms of computing power, storage capacity, data access and visualization. The location of the tremor sources requires robustness and a high quality of data, and continuity of transmission for high number of relevant seismic stations. On the other hand, the use of sophisticated (here unsupervised) classification schemes critically depends on the availability of key stations. Volcanoes, in general, are difficult places from a logistical point of view, making precautions necessary to guarantee a reliable and stable data acquisition. These needs require specific characteristics of the data acquisition hardware regarding the performance of the instrumentation in difficult environmental conditions, problems of power supply and specific precautions for hosting vaults. Since data are transferred preferably digitally via the intranet/internet network, a considerable bandwidth for data transmission is necessary. Up to 100 Mb/s can be transmitted, which guarantees a considerable overhead capacity and allows reliable recording even through time periods with a high signal dynamics (seismic crisis, strong tremor radiation, etc). Under normal, quiescent conditions only 5 to 10% of the total capacity is required for real time data transmission.

In the aftermath of the 2001 and 2002-03 flank eruptions, INGV has undertaken considerable efforts in order to match these goals. Seismic sensors were gradually upgraded installing 40s Nanometrics Trillium sensors, thus having a unified instrumentation available for a large number of stations (33 in the area of Mt Etna).

The data acquired from remote stations regard the whole eastern part of Sicily, where Mt Etna, and the most important Sicilian active volcanoes, are present. The Figure 13 shows a scheme of the data acquisition chain. Data recorded by the various types of sensors are digitized using custom acquisition boards and stored first in a local buffer at the station.

A modern station is essentially a computer with a specific hardware architecture, which guarantees operational stability under unfavorable environmental conditions, and software devoted to certain operations of data handling. In the case of seismic, GPS, accelerometer or infrasound signals, the LibraTM system, manufactured by Nanometrics, Inc. is used. It offers a local data storage which covers about 10 hours of the complete data stream. Each data block stored in the local buffer is marked with a unique ID number.

Data not sent in real time - for instance, because the transmission channels fail for some time - are retransmitted once the connection is restored, and can be easily inserted in their proper place in the data sequence.

¹ The problem is similar to the one met with the identification of nuclear explosions. Waveform data may give important hints and allow the location of the event, the characteristics of it as a nuclear explosion has to be proven by the observation of radio nuclide elements.

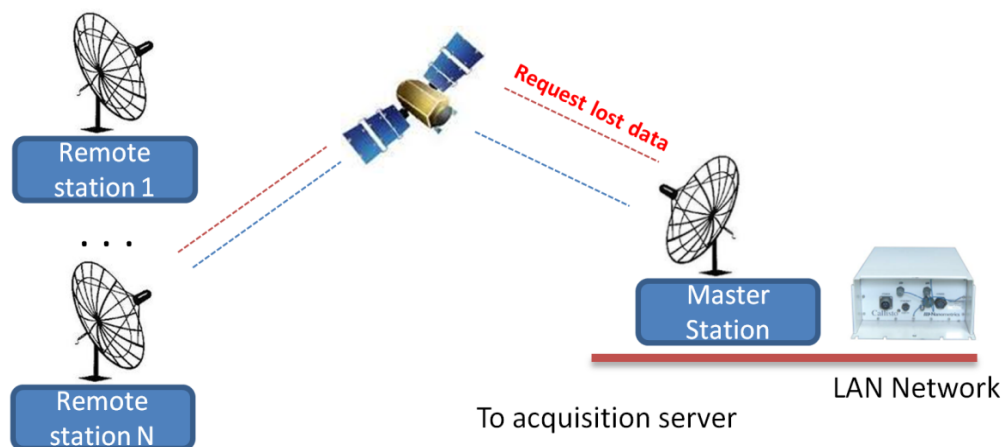


Figure 13. The data acquisition scheme. Data collected at remote stations are transmitted via the satellite base system VSAT to the master station. In the case of seismic data all these units are manufactured by Nanometrics, Inc. and operate under the standards of this company. The Acquisition Server fetches the data from the Master Station and integrates them into the general data management environment of INGV-OE.

In general, data are transferred from the recording site (“station”) to the data centre in real time under various Ethernet transmission protocols, using different carriers (Satellite, UMTS, wireless 5GHz, radio modem). Compared to analog transmission the digital transmission scheme demands a considerable effort both with respect to the technological characteristics of the equipment as well as the necessary bandwidth for data transmission. The advantage relies in the high data quality as well as storage and handling issues (e.g., handling of retransmission requests in case of missing data), which are excluded in an analog acquisition system. Following common standards for modern networks, we apply both TCP-IP network protocols (safe but slower), and, where the risk of error are low, also the faster UDP-IP protocols.

4.3. Handling Incoming Data - The Earthworm Framework

Earthworm offers a global platform for purposes of data centres, addressing issues of acquisition, managing, storing and processing of the recorded data. It is used now worldwide and it is considered one of the most powerful tools for seismic data analysis. The Earthworm platform allows us to handle data of mixed provenience, in the case of INGV-OE those stemming from the Nanometrics system as well as data from still operating older systems with analog data transmission.

Even if Earthworm was originally designed for seismological needs, some of its modules can be applied to other geophysical parameters (e.g., earth deformations, infrasonic signals) or derived data streams (such as volcanic tremor amplitudes, spectra, etc). Being an open source project, the Earthworm community grows every day allowing an easy access to a large variety of contributions. Earthworm has a key role for data acquisition and handling of data acquired through the INGV-OE networks exploiting both already existing and specifically

tailored modules. Earthworm uses a cyclical ring buffer containing all data. Its internal data structure (Winston and derived modules) overcomes short-cuts of the structural limitation of the traditional format like SUDS (Seismic Unified Data System). The SUDS format – as still used at INGV-OE - produces closed files as output, containing all acquired channels in a data stream. As a consequence, data retrieval becomes very tedious as the operator needs to open all channels inside the file even if many of them may not be of interest.

In the Earthworm philosophy each module is continuously watching the data stream in the ring buffer. The modules read messages catching relevant information, perform specific tasks (such as data acquisition, phase picking, etc) and write messages to the ring buffer (Figure 14). This allows for the modules to communicate among each other by broadcasting and receiving various messages via the ring buffer, a principle similar to radio communications (<http://folkworm.ceri.memphis.edu/ew-doc/>). In this way the necessary data traffic is reduced to the amount of information strictly necessary for the operation of the system.

4.3.1. Data Storage - a) The Cluster Architecture

The core of handling and storage is made up of the cluster servers and a storage shelf. A cluster architecture consists of a group (at least two) of independent servers (“nodes”) that work together as a single system (Figure 15). Each node is connected by a physical local network (a local area network, fiber channel connections, etc) and managed by a specific software - called “middleware” - that supervises and monitors the state of health of the whole system. In particular, the middleware catches failure of cluster nodes assigning the processes of the failed node to another one. This so called failover process aims at maintaining all services and applications alive.

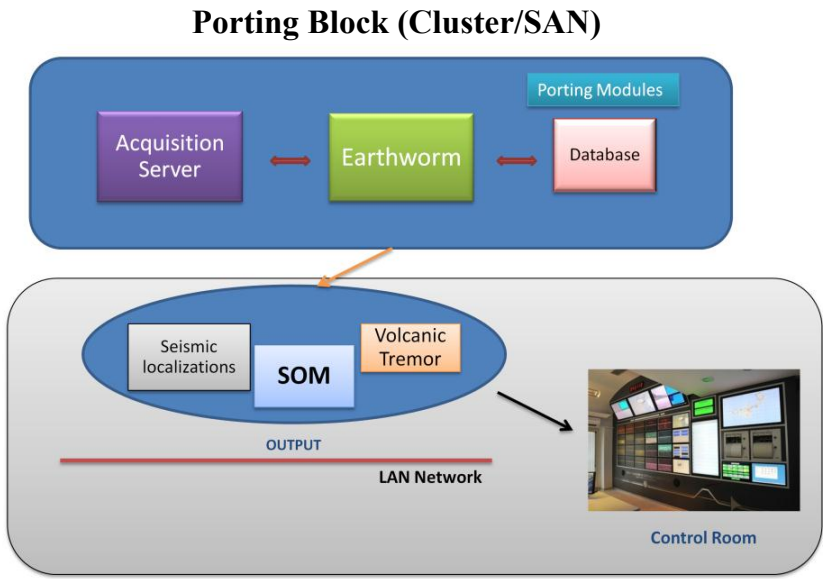


Figure 14. More on Data Storage. Once data are present in the Acquisition Server they are passed to the Earthworm System. Earthworm passes data to the database, and handles newly created data flows, such as the volcanic tremor data, spectra, the SOM/Fuzzy Classifier data. At the end data and results are made available to users and partly displayed in the monitoring center of INGV-OE.

In general, a cluster architecture improves the performance of active systems. In particular it provides:

- i. load balancing: the capacity to share computational workload improving the overall performance. In general cluster architectures bundle a number of low-cost processors together obtaining a high performance computing system. Sharing services, database and computational operations between different nodes, optimizes the performance of real time analysis and reduces the response time for alert messages available inside the INGV-OE Control Room.
- ii. failover architecture: using redundant nodes as previous described in order to minimize/eliminate the single points of failure. This guarantees a high availability of the services.

4.3.2. Data Storage - b) The Storage Shelf (SAN)

A SAN (Storage Area Network) is a collection of storage and backup devices connecting the servers and storage elements through a specific network of channels. Fiber glass channels offer the necessary speed for the transmission of the huge amount of data (about 10 GB/day of seismic data with a total of 400 channels at the moment of writing). They can reach length of up to 10 kilometers, allowing the backup of data in remote sites, which is an additional element of safety.

Compared to conventional storage devices like hard disks etc, a SAN can be considered as a smart data storage infrastructure. It contains a management layer which organizes the data stream inside the storage elements, i.e., disk drives and tapes. One of the most important characteristics of a SAN is that it is easy to reconfigure, for instance adding new storage elements or replacing them meanwhile the system remains fully operative. The SAN in a certain sense forms the final element of the whole acquisition chain.

Its advantages can be summarized as follows:

- i. dynamic allocation of data storage (“Thin provisioning”). The whole storage capacity is available as the physical device of storage is not specified a-priori. Dynamic allocation leads to an efficient management of storage capacities.
- ii. the “snapshot” approach is an effective strategy for disaster recovery. It is based on a differential backup approach monitoring changes / increments of the data structure rather than performing a full copy of all stored data every time. Thus, backups can be carried out frequently and times necessary to recover from failures are limited. Nonetheless it is guaranteed that full data sets can be restored after a disaster.
- iii. “tiered storage”: reduces the economic costs of storage, optimizing the use of disk arrays accounting for different priorities of data access. “High priority” data (i.e., the most used data, data used for early warning, real time earthquake location, visual and thermal imaging, etc) are stored on fiber channel disks (fast but expensive), while older data are moved to a lower tier storage devices, such as the large but relatively slow (and cheaper) SATA disks.

The INGV-OE SAN storage is configured with a large SATA repository (about 8 Tbyte) for all historical data, while current data, which needs quick access, is stored to fast SAS technology disks. At the time of writing our SAN system has a total raw storage capacity of

17.6 TB. Besides a library based on DLT "Ultrium 4" tapes with about 19 TB of raw capacity is used for full backup (Figure 15).

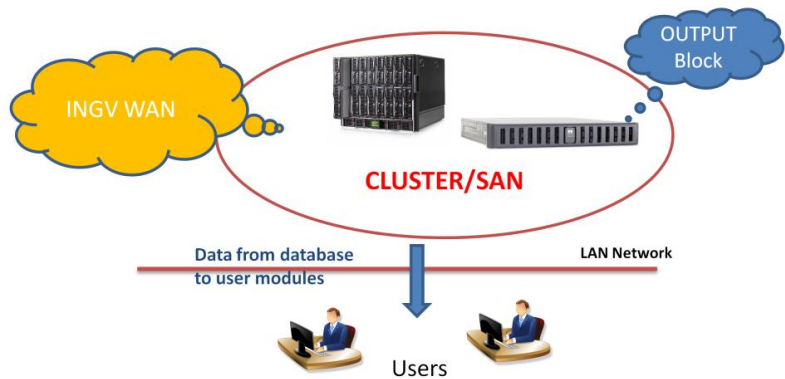


Figure 15. Data management and storage at INGV-OE. The cluster manages issues of both data storage and governs the data access by the INGV-OE users. The SAN represents an intelligent storage architecture warranting both an efficient handling as well as data safety.

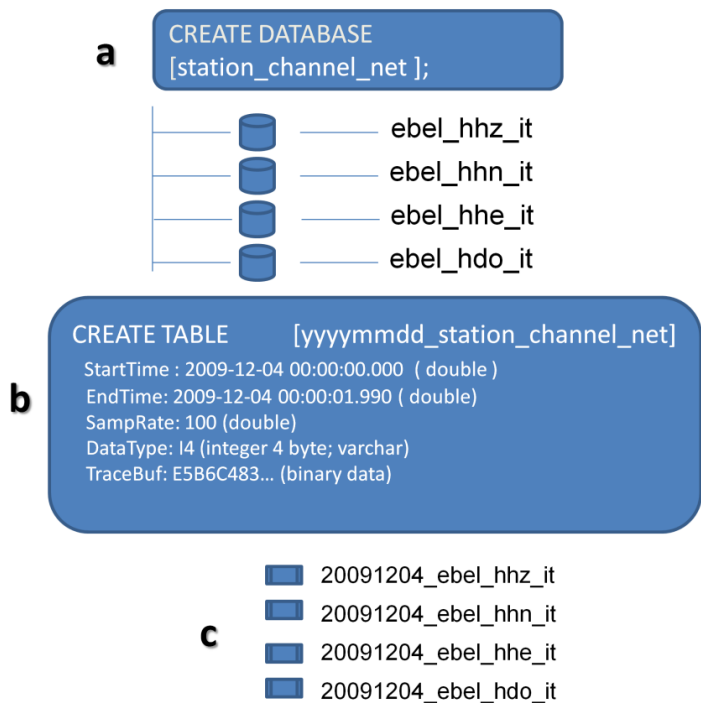


Figure 16. The MySQL database structure flowchart. a) the main script creates the tree structure in the format “ station_channel_net “, here for station "ebel" with three seismic and one infrasound component. For each station the database engine (based on the standard MySQL engine called MyISAM) creates all the tables in the format yyyyymmdd_station_channel_net. b) Description of records in the station table content: start time / end time of the single pattern, data acquisition sample rate (here 100 Hz), DataType description (in our example integer 4 byte length), and “ TraceBuf” field containing raw data in binary format. c) A result of a table created for a record of 2 s on April 4, 2009. Four channels of station “ebel” were selected.

4.3.3. Data Storage - c) Access via Database Concepts

Even though Earthworm has its proper database structure (the “Winston Server”) the variety of information handled at INGV-OE has led software developers to create their own relational database infrastructure. The idea is to allow multidisciplinary access exploiting a relational custom database. The used database is MySQL. Tables are organized (Figure 16) as a tree structure with station names, channels, waveform data, etc. Tables contain short time records, typically one for each second of data acquisition. This allows us to define filters with a high degree of specification which is extremely important for large, multiparametric data sets collected in an observatory like INGV-OE. In fact, the use of data base concepts for handling a-priori selected information, come with significant benefits with respect to memory allocation and efficient use of the available bandwidth.

4.4. Future Developments

In the framework of national projects aimed to technological and infrastructure upgrade, INGV–OE is involved in the development of new strategies for data centres, where complex and multi-disciplinary information have to be administrated. The fundamental principle of this strategy resides in the use of so called “blade servers” in combination with SAN storage technologies. The blade servers strategy follows the idea of increasing the density of computational power in fewer identical servers, in order to reduce management efforts connected with a high number of bundled units of a cluster. A blade server system is housed in a common chassis or rack, hosting many “thin” electronic cards (therefore the term “blade servers”). Each of them forms a complete computer server with one or more CPU processors, memory, network, storage and input/output devices.

Blade servers allow us to host more processing power in less rack space. The hard-wiring is realized inside the chassis avoiding long cable connections in the computer room, at the same time power consumption and heat production is considerably reduced. Contrary to clusters which bundle together a considerable number of low-cost CPU’s, the blade server facilitates the use of multiple virtual machines.

The virtual machines, though being installed on a common server, may even run under different operating systems (e.g, various Windows and Linux operating systems). This reduces the need of software redesign, when new packages developed under differing environments have to be added on. The possibility to duplicate virtual servers allows us to distribute applications improving failure protection, as a failed virtual server can be easily recovered from its virtual image on disks.

Furthermore, in order to assure interoperability between geophysical sites, the use of standard software and database requirements, as well as data presentation and backup, helps the research community in dissemination and data outreach (Figure 17). The basic idea is to improve the use of a Web/GIS portal for the dissemination of results and to share data in a standard scientific format compliant with the most important geophysical/volcanological Observatories around the world. Many collected data will be available inside the Portal respecting some data policy rules. It will be possible to request “on the fly” some shared data: the database procedure will be able to generate some plots and preview tables on request. Furthermore, for the scientific community, tables and packed data will be available for download.

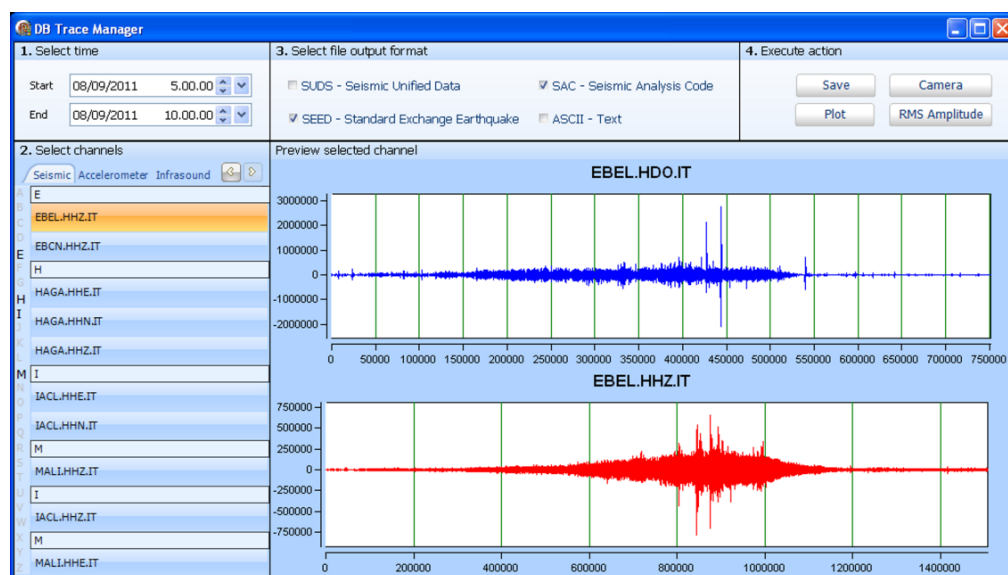


Figure 17. Screenshot of an example of data access. Data can be filtered with respect to stations, channels and time interval of interest. The system offers various output formats in order to facilitate post-processing with external software packages. The buttons on top right stand for a development envisaged in the future. We envisage the options to “save” chosen interval data in a format chosen by the user, “plot” data in a preview window, and to convert the original time series to a sequence of “RMS” amplitudes. Clicking the “Camera” button, a new screen will pop up where the user may select available visual observations (at the moment of writing optic or thermal imaging).

DISCUSSION AND CONCLUSION

Lava fountains as well as flank eruptions are commonly preceded by both increases in tremor amplitude as well as changes with respect to spectra characteristics of this signal. The location of its sources provides further insights about a possible development of volcanic activity on Mt Etna. Eighteen lava fountains occurred in 2011, with eruptive centres close to the South East Crater (Figure 3). In the following we present four representative examples, merging the various observations to a unified picture. We show how the different findings may support each other. The advantage of such a global approach is obvious: an alarm which is supported by more or less independent evidences has always a higher level of significance, as misunderstandings of single parameters can be avoided. At the same time, the use of various criteria provides some redundancy. The “multi station” strategy described in section 2 may find obstacles for instance during winter, with a number of stations not being operating. This affects the quality of tremor source locations, and may pose also problems to the calculation of STA/LTA at a sufficient number of stations. An automatic alert is still possible as long as stations, for which the classifier approach is followed, are alive. At the same time, failure of the key-stations for the classifier is backed as long as the network on the whole maintains a minimum degree of stability and alarms can be issued with reasonable reliability exploiting the distribution of amplitudes across the network.

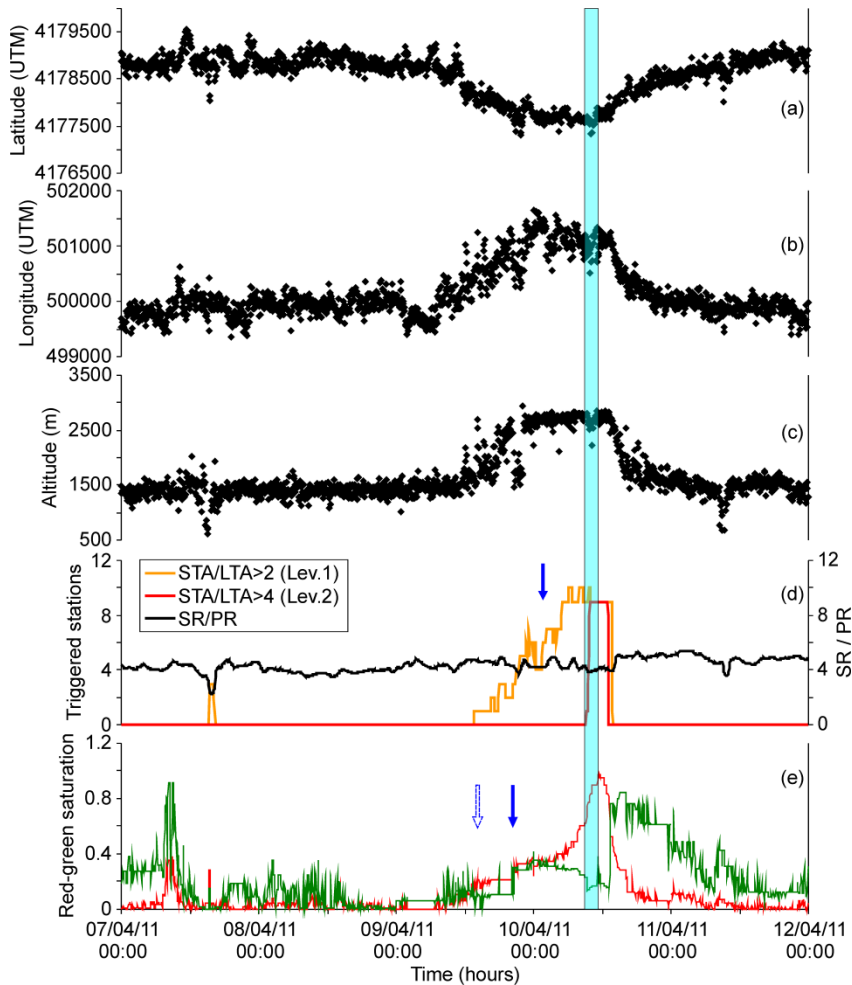


Figure 18. Synopsis of tremor source location (upper three panels). The results of STA/LTA alert system are shown in the 4th panel. The orange line depicts the number of stations with $STA/LTA > 2$, the red line shows the number of stations where $STA/LTA > 4$. Note the black line SR/PR, which reports the precautionary condition discussed in section 2. The classifier based alert system is shown in the bottom panel. The area in turquoise marks the time interval of lava fountain as inferred from volcanological observations (compare Table 1). The solid blue arrows indicate the time when the classifier and the STA/LTA system flag alarms. The dashed blue arrow in the bottom panel shows the flag set by the classifier with the sensitive setting (Table 2a). The time interval, starting on April 7, 2011, includes the lava fountain on April 10, 2011 and lasts until April 11, 2011.

The Figure 18 shows the development of our alert relevant parameters in the time before and during the lava fountain on April 10, 2011. Tremor source locations start to move from a position under the central craters towards the SE-Crater (“SEC”, see Figure 3), already in the morning of April 9, i.e., some 20 hours prior to the unrest becoming manifest at the surface. At the same time, we notice one station flagging STA/LTA criticality (Figure 18d), and the classifier (on station EBEL), using the sensitive settings in Table 2a also flags an anomaly (dashed blue arrow in Figure 18e). In the late evening of 9 April, 2011 the tremor source locations reached a position at ca. 2500 m a.s.l. (that is about 500 m below ground), and at the same time the classifier flags criticality even when using the more conservative settings in

Table 2c (solid blue arrow in Figure 18e). Right before midnight, four stations flag critical STA/LTA and a first level alarm is issued (Figure 18d) and communicated to official authorities. Finally, the unrest becomes manifest at about 9:30 on April 10 , i.e., 10 hours later.

The lava fountain on May 12, 2011 developed more rapidly (Figure 19). In the early afternoon of May 11, tremor sources started moving from their original position, some 2000 m beneath the central craters, towards the SEC approaching more superficial levels. In the sensitive setting (Tab 2a-b) first criticalities were signaled by the classifier at 14:20. At 18:25 the signs of criticality became more evident, finally the first level of alarm was communicated officially at 19:04, when the STA/LTA threshold was passed at four stations. During night, on May 12, at 1:30 the volcanic unrest became manifest. An interesting phenomenon occurred on May 11, when in the early morning hours an earthquake swarm occurred and the STA/LTA threshold was topped at four stations. However, the amplitude relation between summit ring (SR) and peripheral ring stations (PR) was less than three, so the – false - alarm was properly not issued (see section 2, where we mentioned the additional precautionary conditions for alarm with the STA/LTA criterion).

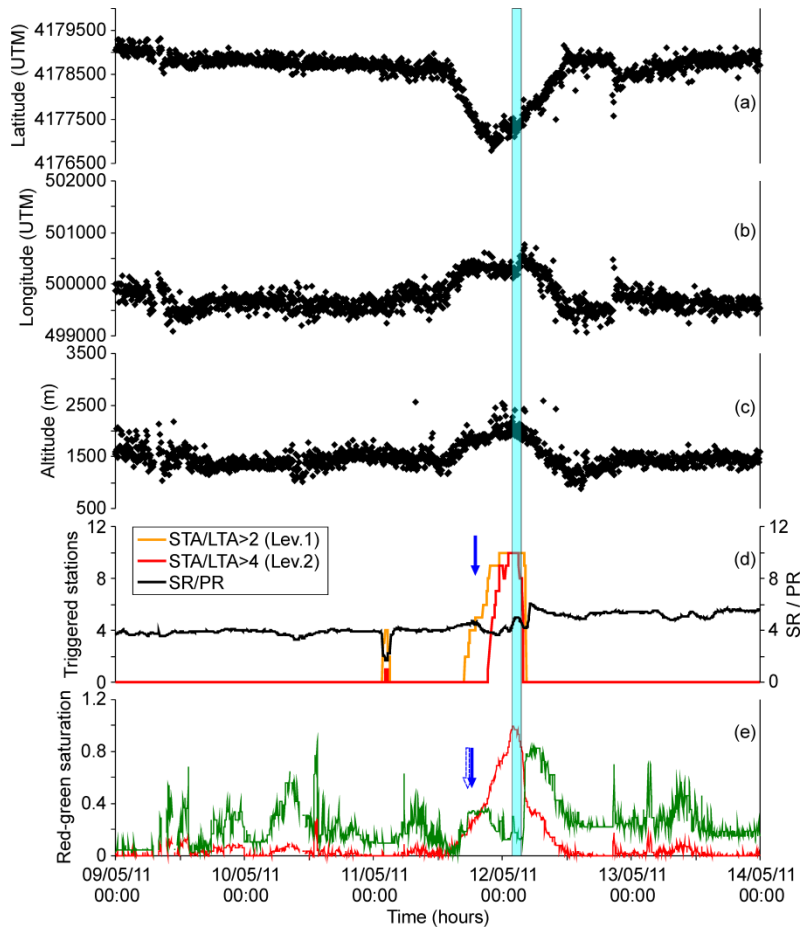


Figure 19. Same as Figure 18 for the time interval from May 9 to May 13, including the lava fountain on May 12, 2011.

The lava fountains on July 30 and September 8, 2011 (Figure 20 and 21) are examples with rather fast development. Tremor source locations shifted rapidly from their original locations under the central area of the volcano towards the SEC, moving upward from ca. 1500 m to ca. 2500 m a.s.l. For the July 30 paroxysm, the classifier signaled criticality at 6:36 and 7:35, depending on its settings, and the STA/LTA threshold (four stations) was topped at 7:35. Lava fountains, however, were noticed only in the evening at 19:30. The September 8, 2011 fountain was a rather short lived phenomenon with a rapid migration of tremor sources. Tremor classifier and STA/LTA alarm systems triggered almost coincidentally, just one hour before strong lava fountains occurred. The short lead time encountered by the various systems is consistent with the rapid movements of tremor sources, which is a clear evidence for rapid development of the phenomenon.

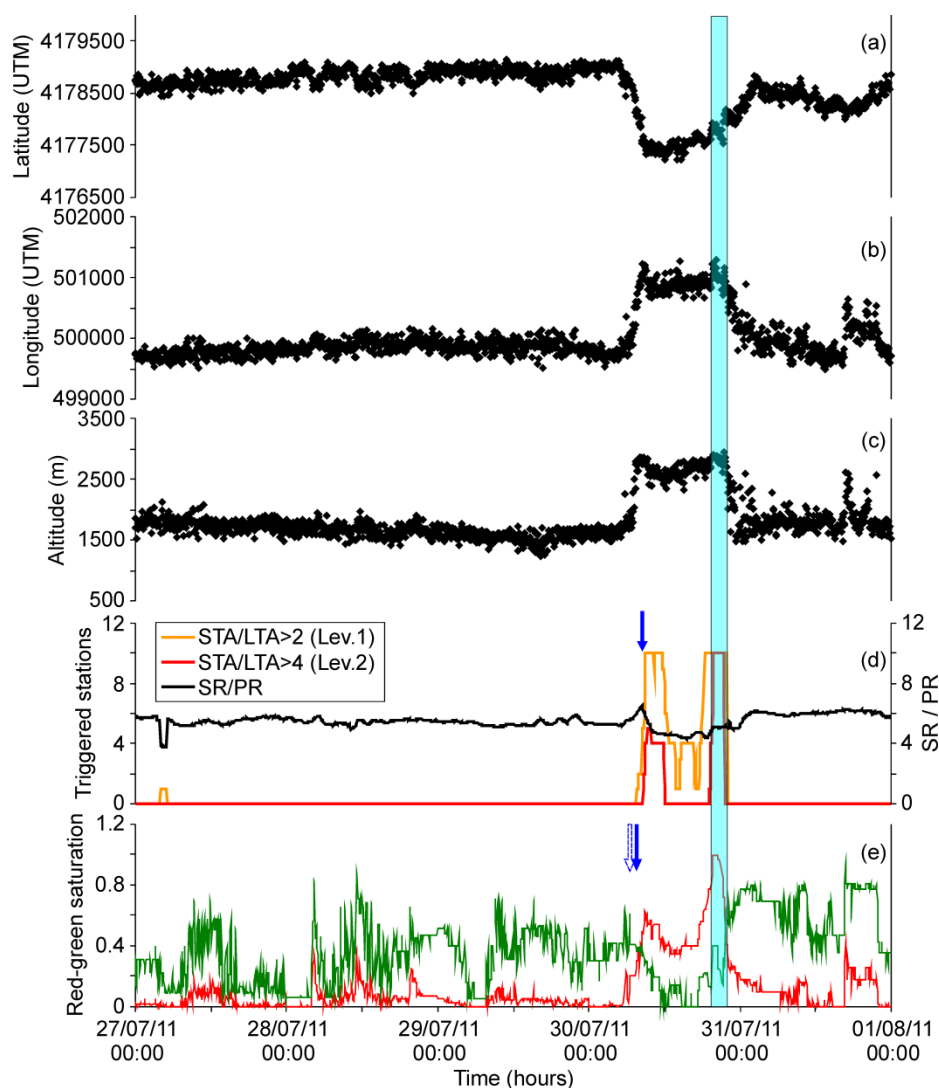


Figure 20. Same as Figure 18 for the time interval from July 27 to July 31, including the lava fountain on July 30, 2011.

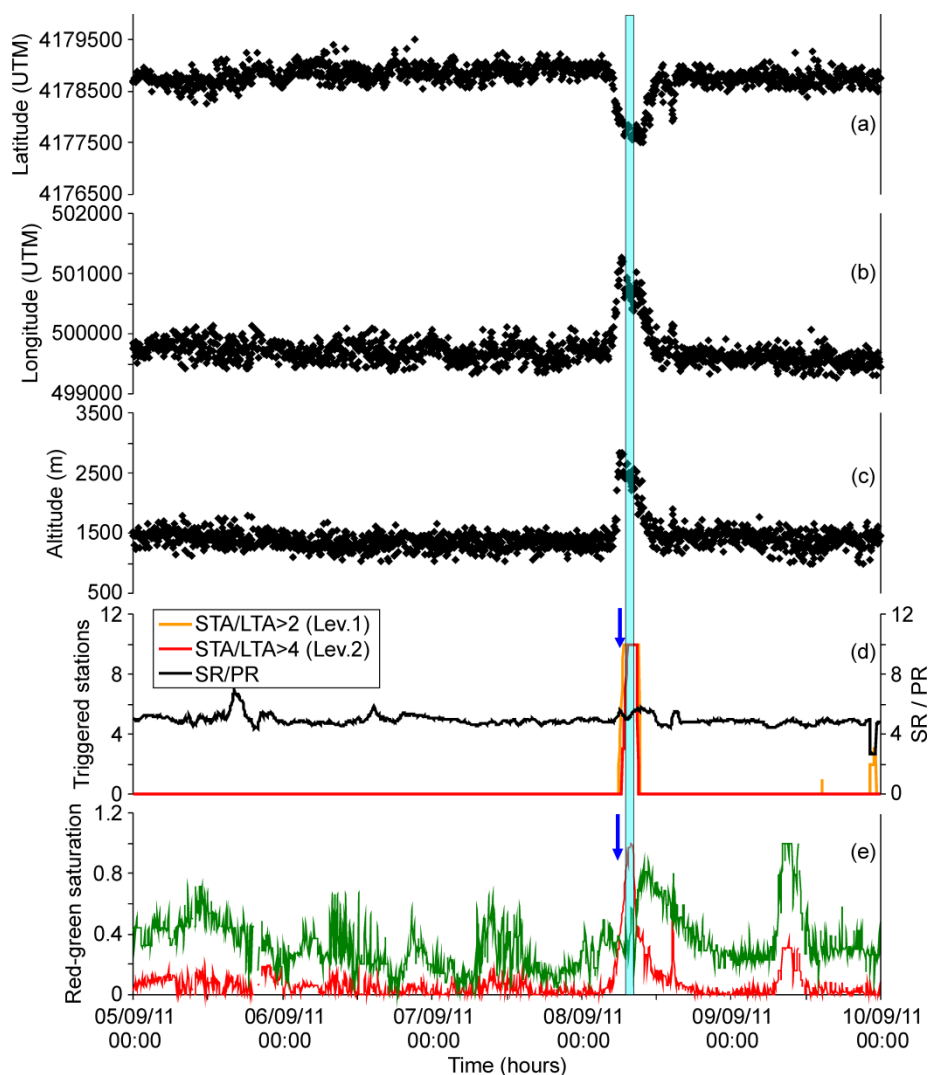


Figure 21. Same as Figure 18 for the time interval from September 5 to September 9, including the lava fountain on September 9, 2001.

Tremor source location, amplitude ratios such as STA/LTA, and the more general characteristics exploited in the signal classifier system yield independent information concerning the state of the volcano. Various protocols for automatic alerts have been developed, all with appreciable results with respect to lead time before the onset of paroxysmal phase of an event, but also concerning reliability – i.e., the low degree of false alarms. On the whole, the classifier approach flags criticality slightly earlier than the STA/LTA criteria. On the other hand it depends on the availability of few key stations, which renders it sensitive to technical problems in data acquisition (power down, interruption of data transmission, etc.). Besides, the system is calibrated for events taking place in the summit or on the higher SE flank of the volcano. Given the lack of experience, we cannot foresee how the system may behave if volcanic activity starts at other areas of the volcano, let them be in other sectors of the mountain or – even worse – at low elevation. The multi station strategy

promises more flexibility in this respect as the seismic network installed on Mt Etna is dense and covers the whole volcanic edifice. Tremor source location is supposed to give reliable results as long as sources do not fall at the very margins of the network or outside. Future developments of the various strategies include a-posteriori applications to older eruptive episodes, such as the ones in 2006. Given the rich material provided by the 2011 lava fountains, the application of the classifier can be extended to more key stations for which the reference material was not sufficient before 2011. This certainly enhances the stability of this approach with respect to failure of a single station.

The technical development of data management, widely exploiting modern data base concepts, will increasingly allow a multi-parametric view of volcanic phenomena on Mt Etna. This concerns other geophysical data, such as infrasound, deformation, magnetic, as well as image based monitoring (e.g., optical and infrared imagery). The comprehensive treatment of various data aims at refining the alert criteria for events like the ones discussed here. Besides, we expect to better focus other critical phenomena, such as ash emissions, which are not necessarily linked to evident seismic signals but are anyway critical for the air traffic. All this information must be managed, safely stored and structured in a way that it can be easily accessed for near real-time applications as well as in a-posteriori analyses in the future.

ACKNOWLEDGMENTS

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APPENDIX

A1. General Remarks

One of the basic steps mentioned in section 3 for reducing the amount of data resides in parameter extraction and data compression. This goal is achieved by defining specific feature vectors – here spectral components - instead of using the original time series. As a consequence we deal with objects rather than with single valued data, which poses particular problems of representation of the characteristics of an object. The choice of the techniques used here is guided by the need to graphically represent the development of these characteristics with time. Both Self Organizing Maps, where the characteristics of objects are expressed by a color, as well as Fuzzy Clustering were found suitable tools for this purpose.

A1.1. Objects, Classes, Patterns, Features

Classification means a process of assigning objects to a category or class. Objects (such as our tremor records) are transformed to patterns which are described by a number of features. Features, in principle can be metrical, ordinal or nominal data. The set of – here

numerical - features make up the feature vector. The choice of the features depends essentially on practical considerations, and is governed by two rules: (i) the features should allow to identify objects uniquely, and (ii) low dimensional feature vectors are preferred to high dimensional ones. Frequently, features cannot be used in their original form but need pre-processing steps, such as the normalization of the numerical values or procedures to limit the dimension of the feature vector. The patterns we are considering here are spectra, and the features are the spectral amplitudes measured in certain frequency intervals.

A1.2. Unsupervised Classification

Whereas supervised classification exploits a-priori knowledge of the class membership of the objects, such an a-priori information is unnecessary in unsupervised classification. The latter concept is based on a suitable definition of similarity between patterns. Similarity is defined on the basis of a distance measure. The task of unsupervised classification can be formulated as to find groups of similar data, i.e., with a minimum degree of heterogeneity. The degree of heterogeneity is defined as a distance measure, such as the Euclidean distance.

A2. Construction of a Self Organizing Map (SOM)

SOM are also known as Kohonen maps after their inventor Teuvo Kohonen (Kohonen, 1984). They are made up of a number of nodes - each one representing a number of patterns - which in a certain sense form a small cluster of their own. Throughout the iterative training process, the node weights are adjusted at every time step so that the sum of the distances between original data and their representing prototype nodes converges to a minimum. Formally, during the training of SOM, one can minimize the sum of the distances with the relationship

$$D_{ij} = \sqrt{(\mathbf{W}_i - \mathbf{V}_j)^T (\mathbf{W}_i - \mathbf{V}_j)} \quad (\text{A1})$$

where $\mathbf{V}_j$ is the normalized input feature vector and $\mathbf{W}_i$ the weights stored in the nodes. $\mathbf{V}$ and $\mathbf{W}$ have the same dimension, i.e., 62 in our case. A core step is the identification of the closest node to the actual input vector, i.e., the best matching unit (BMU) for the j_{th} pattern. Neighbouring nodes lying within a certain radius of influence are considered as well (Figure 5a). Once BMU and the nodes falling within the area of influence are identified, the weights are gradually adjusted according to the so-called learning rate λ , which decreases with time t . A second parameter φ called influence radius describes the dependence on distance Δ of the upgrade of a node. $\varphi(\Delta, t)$ is maximum for the BMU, whereas nodes outside the radius of influence are not upgraded at all. Similar to the learning rate λ , also $\varphi(\Delta, t)$ decreases with time. Various options are available for the shape of φ , such as the Gaussian one, which was adopted in our application. The upgrade of weights follows the relationship

$$\mathbf{W}_i(t+1) = \mathbf{W}_i(t) + \varphi(\Delta, t) \cdot \lambda(t) \cdot D_{ij}(t) \quad (\text{A2})$$

An appealing property of this scheme is that topological relationships among the original data also hold for the BMUs. Consequently, patterns represented by neighboring nodes in the SOM are truly close to each other also in the original data space. This characteristic of topological fidelity is indeed achieved by using the influence radius $\varphi(\Delta, t)$. An in-depth description of SOM is reported in the most recent edition of Kohonen's classical work (Kohonen, 2001).

An intuitive example for the representation of pattern characteristics with SOM is the so-called World Poverty Map, created on the basis of 39 socio-economic parameters collected worldwide (Figure 5b). In our application, we use an analogous approach for our patterns where we have 62-dimensional feature vectors. In Figure 5c we schematically represent the neighborhood relations of the SOM nodes for our data set. We chose to use to represent the nodes as hexagons. Numbers marked in the hexagons indicate for how many patterns each node is the best matching unit. For the definition of the Topological error (see below) we shall also consider nodes which are the second closest ones to the patterns. Some nodes never happen to be a BMU; in this case, their corresponding hexagons are marked with zeros.

We designed the map following Vesanto et al. (2000), who suggested that the number of nodes should be fixed by taking ca. five times the square root of the number of patterns (about 4400 for EBEL station) – i.e., ca. 330 nodes. The aspect ratio of the map depends on the square root of the ratio between the two eigenvalues of the covariance matrix, which is ~ 7 in our case. With $7 \times 49 = 343$ nodes, the map shown in Figure 5c matches both the desired number of nodes and the aspect ratio with a good approximation. Figure 5d gives an approximate picture of the true distances between the nodes of the map. The position of each node is represented in a system of axes made up by the eigenvectors corresponding to the two largest eigenvalues.

As SOM can be particularly instructive when the weights take the form of a color code, we apply Principal Component Analysis, transforming the weights in a 2D representation space made up of the principal axes z_1 and z_2 . These axes are normalized in the range $[0 \dots 1]$. A third value z_3 is obtained either as $1 - z_1$ or $1 - z_2$. We assign the principal colors: red, green, and blue to each of the z_i . Each node receives a mixture of colors according to its original weight. The color coding allows us to visualize the time dependent behavior of tremor characteristics by a sequence of colored symbols.

A3. Fuzzy Clustering

In cluster analysis we follow a non-hierarchical, partitioning strategy of clustering. The measure of the heterogeneity of the k_{th} cluster is expressed as the sum of the squared distances of the patterns in each cluster from their centroid vector. In crisp clustering each pattern is exclusively assigned to one cluster (Figure 6a). Thus information about pattern characteristics shared by two or more classes is lost, as the attribution to a cluster is unique.

A solution to this drawback is fuzzy clustering, in which each pattern may belong to a certain degree to all possible classes. Consequently, the class membership of a pattern is given by a vector. The partition is described by a $M \times K$ matrix of class membership values, with M being the number of patterns in the entire data set, and K the number of clusters. In fuzzy clustering the optimum partition is determined minimizing the so-called Objective Function J

$$J = \sum_k J_k \quad (\text{A3a})$$

$$J_k = \sum_i m_{ik}^q d_{ik}^2 \quad (\text{A3b})$$

where m_{ik} is the membership value of pattern i with respect to cluster k , d_{ik} is its Euclidean distance from the cluster centroid, and q is a weighting exponent (set equal to 2 in our application). The cluster centroid is obtained from the relation

$$\mathbf{c}_k = \sum_i m_{ik}^q \mathbf{u}_i / \sum_i m_{ik}^q \quad (\text{A4})$$

$\mathbf{u}_i$ being the i_{th} pattern vector. The components of this vector are given by the membership values m_{ik} in eq. (A3b), and express the degree to which a pattern belongs to a given cluster (Figure 6b). The larger the number of those patterns, the closer the partition to crisp clustering.

In non-hierarchical partitioning cluster analysis the number of desired clusters has to be chosen a-priori. As a basic rule, one tries to fix as few clusters as possible, having the highest degree of non-fuzziness. We chose three clusters as ~75% of the patterns showed a prevailing membership degree of 0.7 or higher. With more clusters the number of patterns with a high prevailing membership degree decreased. This result comes from the fact that the higher the number of clusters, the larger the number of centroids competing for the patterns.

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Chapter 5

FROM THE STANDARD STROMBOLIAN ACTIVITY TO THE VOLCANIC CRISES AT STROMBOLI (ITALY): AN OVERVIEW OF THE SEISMIC SIGNATURES

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ABSTRACT

The properties of the seismicity of Stromboli volcano preceding and during the 2002-2003 and 2007 crises have been reviewed on a statistical basis. Both the crises show the response to the rising magma in the seismic observations but on very different time scales. A statistical analysis on all explosions reveals that the occurrence is always driven by a Poisson process in both crises even when an effusion phase occurs, with the only difference in shortening the inter-times just during the effusion. A slightly different process can be advocated for the swarms which display a unimodal distribution. Regarding the amplitudes of the explosion-quakes, they have a log-normal distribution until the effusion onset as in the standard Strombolian activity (stationary phase). The departure from the stationarity to a non-stationary phase seems to be traced by a precursor evidenced in the tidal regime, which is related with the duration of the crises. It appears as a transient oscillating signal at about 3-days that modulates the explosion amplitudes in 2007 and as a variation in the inter-occurrence among explosions in the longer crisis of 2002-2003. The successive activity can be interpreted as the response of volcano to restore the equilibrium condition. The escape from the equilibrium produces, at first, variations in the usual statistics of the explosions, then it leads to the lava effusion and to a pressure drop in the plumbing system that induces a deep gas slug nucleation and the excitation of low frequencies.

Keywords: Strombolian activity; Poissonian Process; Log-normal amplitude distribution; modelling

INTRODUCTION

Stromboli is a strato-volcano located in Tyrrhenian sea on the homonymous island at North of the Sicilian coasts (Italy). Its volcanic activity consists of explosion-quakes occurring every 5-10 minutes that eject gas and scorie superimposed to a continuous persistent volcanic tremor. Researchers refer to that activity as "Strombolian" also when occurring on other volcanoes [see, e.g., Erebus volcano (Antarctica), Aster et al., 2003; De Lauro et al., 2009a,c; Etna volcano, Patanè et al., 2008; Popocatepetl, Arciniega-Ceballos, 1999]. Stromboli cyclically escapes from its stationary standard behaviour to experience a non-stationary stage. From a seismological point of view, the stationary behaviour is characterized by both a continuous tremor normally distributed in amplitude and low-intensity explosions with a log-normal amplitude distribution and Poissonian inter-occurrence times [Bottiglieri et al., 2005; De Lauro et al., 2008]. During the non-stationary phase, effusive and paroxysmal events may occur, which can likely trigger landslides and tsunami.

Recently, the stationary activity has been interrupted in 2002-2003 and 2007 by eruptive crises with episodes of more vigorous activity accompanied also by lava flows. The study of how this volcano escapes and goes back to the stationary regime is of great interest from a volcanological point of view and for the physics of fluid systems as well; indeed, the physics of the non-equilibrium systems and the consequent turbulent behaviours can be considered one of the major challenge for theoretical and experimental researchers in many fields of science and technology. Obviously, it is necessary to study the phenomenon over the whole process, i.e. starting from the escape from the equilibrium, through the development of the eruptive crisis, up to the return to the stationary state. In the case of Stromboli it is possible to follow the entire process, because it has been monitoring for the last 30 years with permanent instruments and temporary networks. Here, we review the seismic features during the transition from the standard Strombolian activity to the eruptive crises of 2002-2003 and 2007. During 2002 crisis, the continuous seismic monitoring started with the deployment of a permanent digital broadband network [Martini et al., 2008]. A description of the seismic activity at Stromboli before the 2002-2003 eruption is based on data provided by temporary seismic surveys [see, e.g., Dietel et al., 1994; La Rocca et al., 2000; Chouet et al., 2003; Pino et al., 2004; De Lauro et al., 2008]. On the contrary, for the 2007 we take advantage of a three-component broad band station and of a borehole strainmeter for a some limited period.

Here, we review the spectral properties, the energy and the inter-times of the Strombolian explosion-quakes during the equilibrium-nonequilibrium transition. The original data-set were composed of tens of thousands of events (see Section 3) which allows to detect the changes of the explosive process during the transition in a statical sense. This approach examines the two time scales involved: the duration of the explosion-quakes (i.e., tens of seconds) for the waveform analysis and a longer scale (from hours to months).

STANDARD STROMBOLIAN ACTIVITY

We remind that standard Strombolian activity is made of explosions superposed to a persistent background volcanic tremor. The latter is characterized by a broadband spectrum in

the range 0.1-7 Hz. Decomposition in time domain by means of an entropy-based technique [De Lauro et al., 2005a and references therein], i.e. Independent Component Analysis (ICA), shows that tremor is a linear superposition of four signals that can be recognized as self-sustained oscillations [De Lauro et al., 2008]. These signals are characterized by frequencies in specific ratios and show a clear radial polarization pattern with a dominance of P-waves [see, e.g., Falsaperla et al., 1998; De Lauro et al., 2008]. The lowest band (0.1-0.5 Hz) with a characteristic peak close to 0.3 Hz contains the best polarized signal, sometimes masked by contributes coming from oceanic microseismic noise [De Lauro et al., 2005b, 2006], whereas scattering prevails at higher frequencies. The explosion-quakes are an enhancement in amplitude of the volcanic tremor [De Martino et al., 2002; Ciaramella et al., 2006] induced by the formation, ascent and bursting of gas slugs. As well as tremor, they can be decomposed into well characterized seismic signals [Acernese et al., 2003] displaying a radial polarization pointing to the crater area [Acernese et al., 2004].

The inter-occurrence times between two successive explosion-quakes (inter-time) is a measure of the waiting time between successive slugs. It is driven by a Poissonian process, whose rate is related to the apparent repetitiveness of the phenomenon [Bottiglieri et al., 2005], implying absence of interaction among the bubbles. The stationary phase of Stromboli including both tremor and explosions is well reproduced at macroscopic level in terms of low turbulent systems displaying intermittence [Venkataramani et al., 1996; De Lauro et al., 2009a]. Indeed, the system is generally quiescent (background volcanic tremor) with sudden bursts of activity (explosion-quakes). The switching between bursts and quiescent states occurs following a Poissonian process.

In this unified scheme, both tremor and explosions share a common source to be sought in the thermodynamic state of the volcano. As it can be simply recognized, the thermodynamical conditions of the magmatic system of Stromboli imply that at suitable depths (0.7-2 Km) magma must degas. The exsolved gas, driven by a diffusive dynamics, hits the wall of the volcanic conduit.

By the coupling between the gas flux and the rock, volcanic tremor is produced as self-sustained oscillations. Indeed, confined jets exhibit well-organized oscillation patterns contrary to the convective instability of a free jet and they are the result of a feedback effect [Maurel et al., 1996]. The gas particles, due to the low diffusion coefficient, are confined in a very thin volume (with respect to the whole depth of volcano conduit), favoring the coalescence and then the formation of the slugs [Chandrasekar, 1943; Binder and Stauffer, 1976; De Lauro et al., 2009a].

The Chandrasekar-Landau model is adequate to describe coalescence during an equilibrium (stationary) phase. Once a suitable size is reached (fixed by the conditions of equilibrium), the slugs rise along the volcanic conduit very rapidly, exploding then at the air-free surface. They are expected to oscillate with the same frequency of the background signal reinforcing it both to form and to remain coherent during the rising. This model well describes the transition from tremor to explosion at a coarse-grained level. A detailed quantitative description of the vibrations would require two coupled sets of field equations: one for the solid and another for the fluid oscillations, for a preliminary model see [Lane et al., 2001]. Actually, in the framework of the theory of the dynamical systems, the problem can be simplified and the ground vibrations can be reproduced in time and frequency domains by very simple systems of nonlinear oscillators in the regime of self-oscillations [Konstantinou, 2002; De Lauro et al., 2008; De Lauro et al., 2009b].

DATA-SET

The data-set consists of the recordings of seismic stations operating in the two selected time periods (2002-2003 and 2007). Regarding the first period, which spans from 23/05/2002 - 30/01/2003 (TP1), the only available station (labeled SX15) was equipped with a three-component Lennartz 3D-5s seismometer (LE-3D/5s) having flat frequency response in the range 0.2-40 Hz and a 24 bit RefTek 72A07 digitizer with sensitivity of 400 V/m/s [De Martino et al., 2011a]. The sampling rate was equal to 50 sps [Pino et al., 2004; Cimini et al., 2006]. The station was located about 2.5 km far from the craters in NE direction.

In Figure 1 we show the map of Stromboli emphasizing the positions of the sensor and the craters. The recordings include the first 34 days of the crisis: effusion began on December 28th, 2002, when a fissure opened at the base of the NE crater, and lasted until July 6th, 2003 [see e.g., Calvari et al., 2005; Cesca et al., 2007].

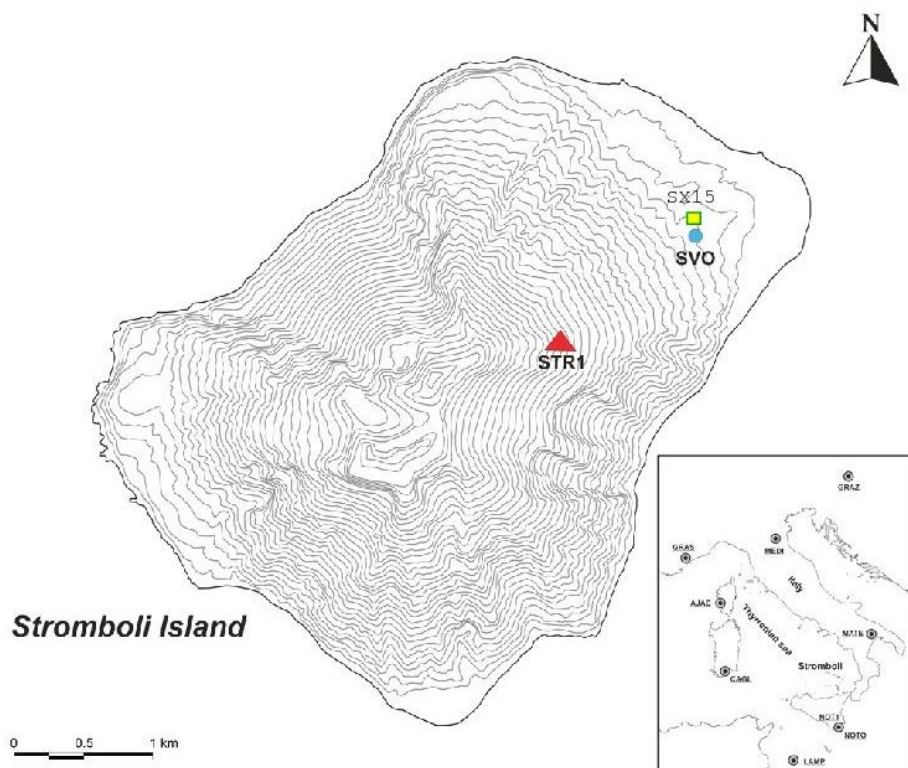


Figure 1. Map of location of both seismometers labeled SX15 (working during TP1), STR1 and strainmeter labeled SVO STR1 (both working during TP2).

On December 30th, 2002, portions of the NW flank collapsed inducing two landslides in the area of Sciara del Fuoco which, in turn, produced a tsunami. On April 5th, 2003, a paroxysmal explosion occurred in the crater area, the largest one since 1930. The effusion was mainly fed by a fissure at about 550 m a.s.l. along the Sciara del Fuoco until mid-February 2003, and from a fissure formed 170 m above since then and until the end of the crisis. We remark that during 2002-2003 crisis, only a seismic station was operating for a

limited time period, namely from 23/05/2002 to 30/01/2003, then including the onset of the crisis and the effusive phase, but excluding the paroxysmal.

Relatively to 2007, the data-set is composed of seismic data acquired by STR1 station (equipped with a Guralp CMG-40T three-component broadband seismometer) in the period 07/10/2006 - 31/03/2007 (TP2) [De Martino et al., 2011b].

Moreover, two strainmeters managed by the "Analysis and Monitoring of Environmental Risk" center (AMRA) have been recording since 2006 at Stromboli, the first at the Timpone del Fuoco site (TDF), close to the small village of Ginostra, and the second close to the local Civil Defense Centre (SVO).

For our aim, we use the latter because it is better coupled with the surrounding rock after the period of stabilization and spanning the period 20/01/2007 - 30/06/2007. Notice that throughout the text, we will indicate as effusive phase (EP) the period 2007/02/27 - 2007/04/02 and as non-effusive phase (NEP) the period before (see Figure 1). We frequency filter the data to guarantee the linear response of the seismometer adopting a high-pass filter with four poles with a cut-off at 0.2 Hz.

SPECTRAL ANALYSIS

The common source process underlying the generation of both tremor and explosion-quakes relative to the stationary phase is reflected into a similar spectral content of the seismic vibrations. The basic properties are preserved during a non-stationary activity such as the effusive phase.

In Figure 2 a normalized spectrogram relative to 2002 is reported. Each bin on the time-scale represents the averaged normalized spectra of 300 successive explosions. It shows that the highest spectral peaks are concentrated in bands, roughly centered at 1 Hz, 2 Hz, 2.5-4 Hz, respectively.

During some periods, other bands around 0.3 Hz and 5 Hz can be distinguished, as well. From the pattern of Figure 2, the authors recognize two principal variations of the basic spectral content (Phase B).

The first variation occurs approximately in the period 20/06/2002 - 05/08/2002 when the band at 5 Hz appears and, at the same time, the band at 2 Hz disappears (Phase A). Another spectral change occurs at the onset of the effusion (28/12/2002) and lasts until the end of the recordings, namely the lowest spectral peak (0.3 Hz) has an abrupt increase in amplitude and consequently the overall shape of the spectrum during the effusive phase has an envelope with a monotonic decrease (Phase C).

They also note two clusters of quasi-monochromatic explosions in the period 13-26/11/2002 highlighted by black boxes of Figure 2 (Phase D). Ultimately, they recognize four classes of spectra reported in Figure 3. Before the effusion, the spectra are bell-shaped with maxima mainly falling between 2 and 3 Hz.

In this phase, the amplitudes of the frequencies around 2 Hz may be low (Phase A) or high (Phase B) and the principal peaks are located at 1, 3, 5 Hz (Figure 3a) and at 1, 2, 3 Hz (Figure 3b), respectively. After the onset of the effusion (Phase C), the envelope of the spectra becomes monotonically decreasing with the fundamental peak at 0.3 Hz (Figure 3c). The clusters of explosions (Phase D) have their predominant peak at 2.8 Hz (Figure 3d).

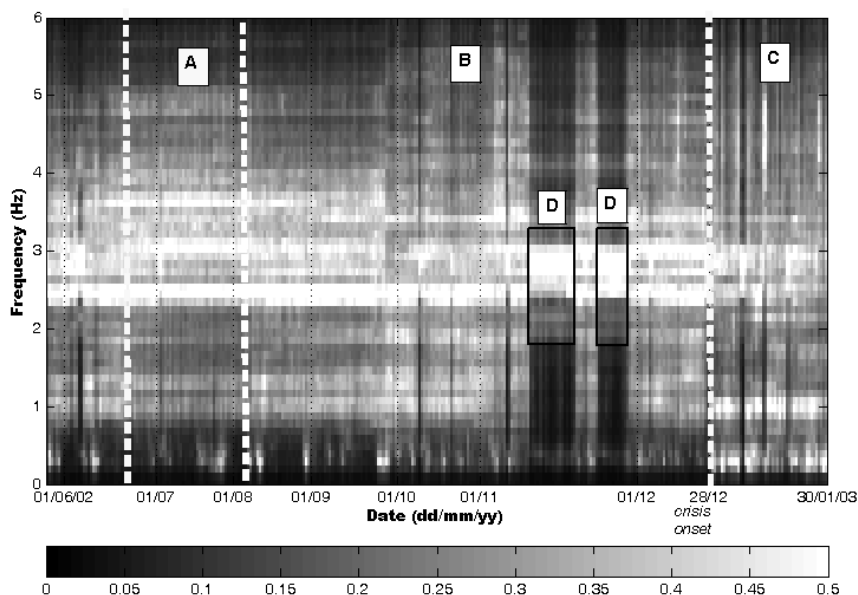


Figure 2. Time evolution of the spectral content of the normalized amplitude spectra of the explosions. The color scale is chosen to optimize the visualization of the spectral changes. The timescale is not uniform because the occurrence rate of the events is not constant. Four different spectral classes labeled as phases A, B, C, and D can be distinguished. Phase B includes the basic spectral content, showing principal peaks at about 1, 2, and 3 Hz. From the end of September 2002 until the crisis onset, these principal spectral peaks increase in amplitude, and high-frequency modes (4, 5 Hz) emerge, suggesting that an energy enhancement is occurring. Modified after De Martino et al. (2011a).

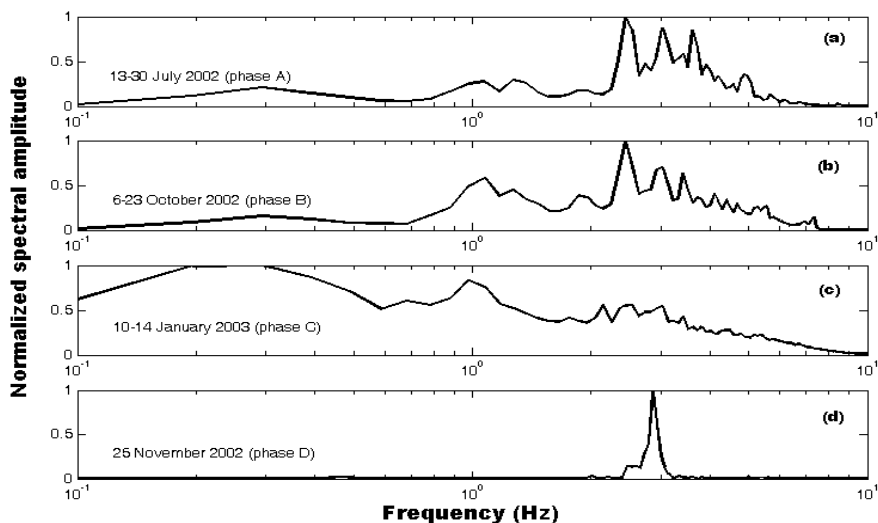


Figure 3. Mean normalized amplitude spectra of four representative samples of signal calculated as the mean spectra of the explosions occurring in the periods (a) 13-30 July 2002 (phase A); (b) 6-23 October 2002 (phase B); (c) 10-14 January 2003 (phase C, which is during the crisis); and (d) 25 November 2002 (phase D).

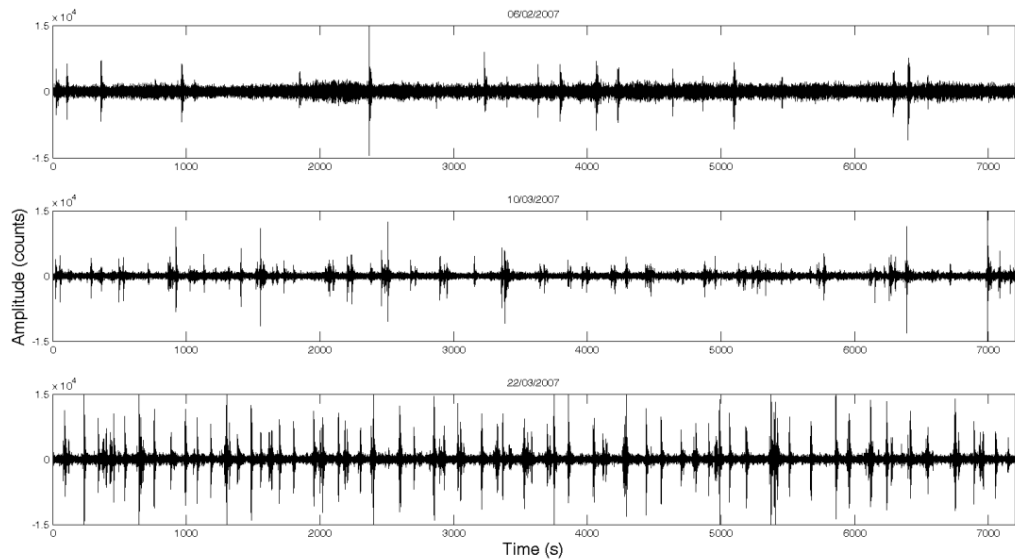


Figure 4. Examples of waveforms recorded during NEP and EP.

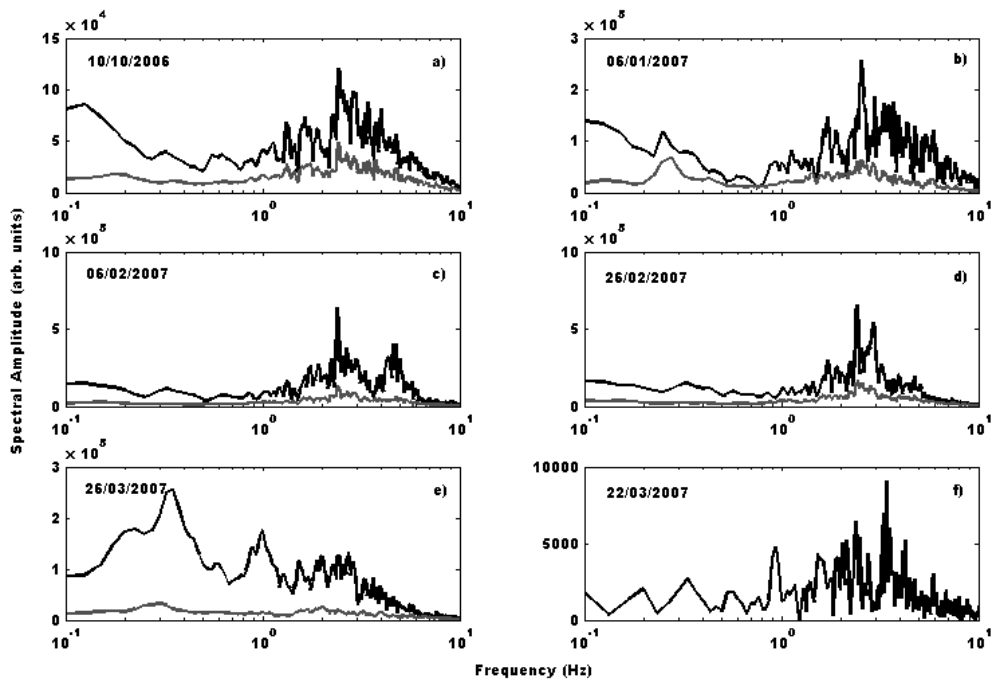


Figure 5. Spectral amplitudes of (a-e) tremor (gray), (a-e) explosion-quakes (black) and (f) swarms acquired in different periods indicated in the legend. The graphs have been computed by averaging the spectral content of tremor and explosions occurring within 40 minutes. The number of analyzed explosions ranged between 4 and 20, according with the variable occurrence rate. Note also the increasing amplitude of the explosions moving towards the effusion onset. The spectrum of the swarm shows a clear shift towards higher frequencies with a dominant spectral peak at 3.5 Hz (f).

Regarding 2007, some examples of waveforms relative to NEP and EP are plotted in Figure 4, in which an increase of the number of explosions moving towards EP is detectable.

In Figure 5, the spectra of tremor (gray), explosion-quakes (black) and swarms (black) are shown. The spectra of tremor and explosions during NEP are clearly comparable, as expected for the standard Strombolian activity: they show standard bell-shape and main peaks in the range [1-5] Hz with the most energetic centered at about 2.5 Hz.

Further lower frequencies can be easily recognized in the spectra: the standard 10s-signal related to the envelope of explosions, and the 3s-signal related to the volcanic tremor [De Lauro et al., 2005b]. Spectral changes occur during EP: the amplitudes of low frequency peaks ($< 1\text{Hz}$) increase so that 0.3 Hz becomes dominant as reported in Figure 5e, which shows an individual case of tremor and explosion recorded on 26 March. This suggests a change in the vibrating structure that can involve a different geometry and/or size or depth of the plumbing system. Finally, the spectrum of an explosion in the swarm shows a clear shift towards higher frequencies with a dominant spectral peak at 3.5 Hz (see Figure 5f).

In conclusion, the behaviours of both TPs can be considered very similar. In a suggestive and clarifying picture, the phenomenology can be cast in a scheme of an overflow mechanism in a cavity. This justifies the enhancement in amplitude of the fundamental frequency band until to its disappearance with the second harmonic that becomes more and more relevant. The system has been experiencing a change of its thermodynamical conditions induced by an increased pressure. The only difference is that the volcano attempts to react developing a more complex behaviour than that in the stationary phase. To have a major insight, the analysis on other time scales is needed.

INTER-TIME AND ENERGY ANALYSIS

Previous sections establish the seismic wavefield properties on the time scale of the explosions' duration. To get an overall understanding of the explosion generating process, it is worthwhile to study the macroscopic behavior of the volcano on a longer time scale, i.e., investigating the distribution of inter-times. Indeed, the distribution function of the inter-times contains significant information on the dynamic process generating the bursting gas slugs. A careful observation may shed light on significant fluctuations in the source mechanism providing information on the transition between stationary and non stationary phases.

Regarding TP1, a decreasing exponential function well fits the inter-time distributions also performing the analysis on different time windows in the range 3-150 hours (De Martino et al., 2011a). It indicates an underlying Poissonian process as detailed already shown in Bottiglieri et al. [2005] and De Lauro et al. [2008]. This is also confirmed by the estimation of the variability coefficient (C_v). We remind that this value is defined as the ratio between the standard deviation and the mean value. $C_v = 0$ when the process is periodic, $C_v = 1$ for a Poisson process, and C_v for a clustered process [Cox and Lewis, 1966]. Figure 6 shows an example of time evolution of C_v estimated in windows of 8 hours: 94 % of the points are in the range 0.7 - 1.3 and 6 % are well below and fall around 0.5. Moreover, in Figure 6, the time-evolution of the occurrence rate estimated as the inverse of the mean inter-time in sliding window of 12 hours is also reported. Occurrence rate shows fluctuations on many time scales around a mean value of about 0.004 s^{-1} , that corresponds to a mean inter-time of 250s. Moreover, some oscillations in the occurrence rate with a periodicity of 3-5 days in the time-interval 15/09/02 - 05/10/02 can be evidenced. The occurrence times are well described

by a Poisson process both before and after the crisis onset, except for some clustered events i.e., swarms that follow different statistics. With respect TP2, Figure 7a contains the inter-time distributions in a semi-log scale; this function does not change its exponential form, thus allowing the estimate of a characteristic rate as best-fit. The different slopes (Figure 7b) are related to different rates, ranging within $[0.001-0.016] \text{ s}^{-1}$. Highest rates are relative to the time-windows during EP. The authors compare the inverse of the mean values of the inter-times with the estimated rates: the time-evolutions of both the parameters are very similar (Figure 7c). This is a further indication that the mechanism driving the occurrence of the explosions is well described by a Poissonian process.

Moreover, the most significant change is located at the onset of EP, where the rates grow considerably. Indeed, during EP the occurrence rates are twice the rates of the NEP standard activity, i.e., from 0.004 to 0.009 s^{-1} . The Poissonian process description is once again supported by Cv; indeed according with the curves of Figures 8a,b, Cv is meanly centered at 0.8 (De Martino et al., 2011b). Cv lowers to 0.6 in correspondence to the swarms, indicating that they are ruled by an eventual different process, which is represented by the distributions of Figures 8c,d. The shape is no more exponential and a maximum appears indicating a most frequent inter-time of about 30 s .

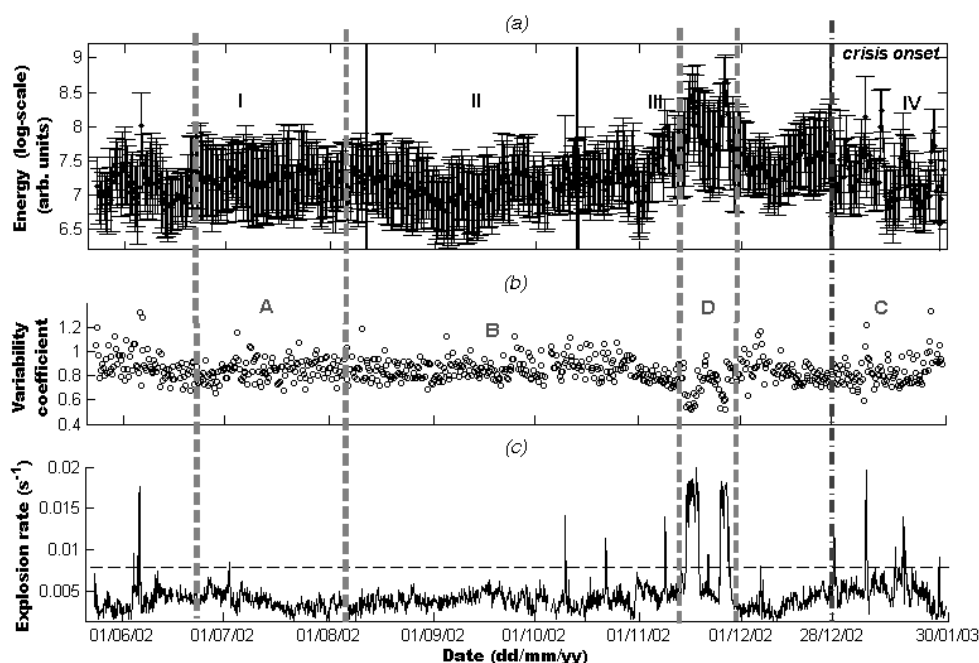


Figure 6.a. Time evolution of mean value and standard deviation (plotted as error bars) of energy computed in nonoverlapping time windows of 12 h . Energy is in log scale. The black dotted vertical line indicates the crisis onset. The three black lines (a) separate the four time intervals cited in the text, labeled as phase I, II, III, and IV. The dashed gray lines represent the edges of phases A, B, C, and D previously introduced on the basis of the spectral content. The time interval of the phase C coincides with that of the phase IV. (b) Time evolution of the variability coefficient estimated in sliding and nonoverlapping windows of 8 h . (c) Occurrence rate estimated as the inverse of the mean intertime in nonoverlapping windows of 3 h . Swarms are detected when the occurrence rate exceeds 1 (represented by the dashed horizontal line). Modified after De Martino et al. (2011a).

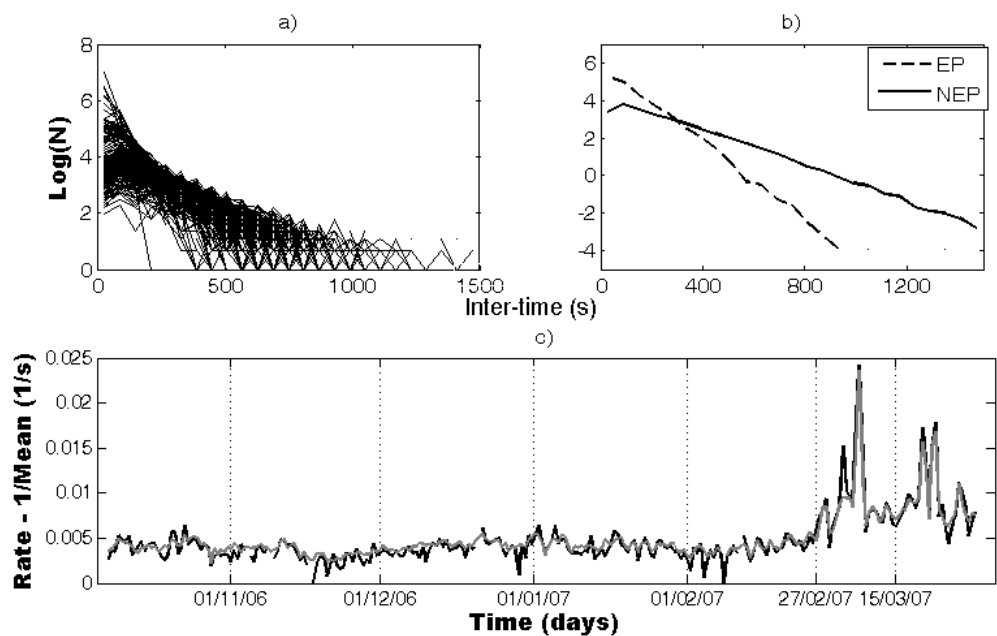


Figure 7.a. Distributions of the inter-times plotted in semi-log scale. Each curve is relative to the events occurring within 16 hours time window; (b) Inter-time distribution averaged over NEP and EP. The change in the slope of the distribution is evident; (c) Reciprocal of the mean values of the inter-times (black line) and rates (gray line) of the explosive process. The rate is computed as the best fit of the exponential decrease.

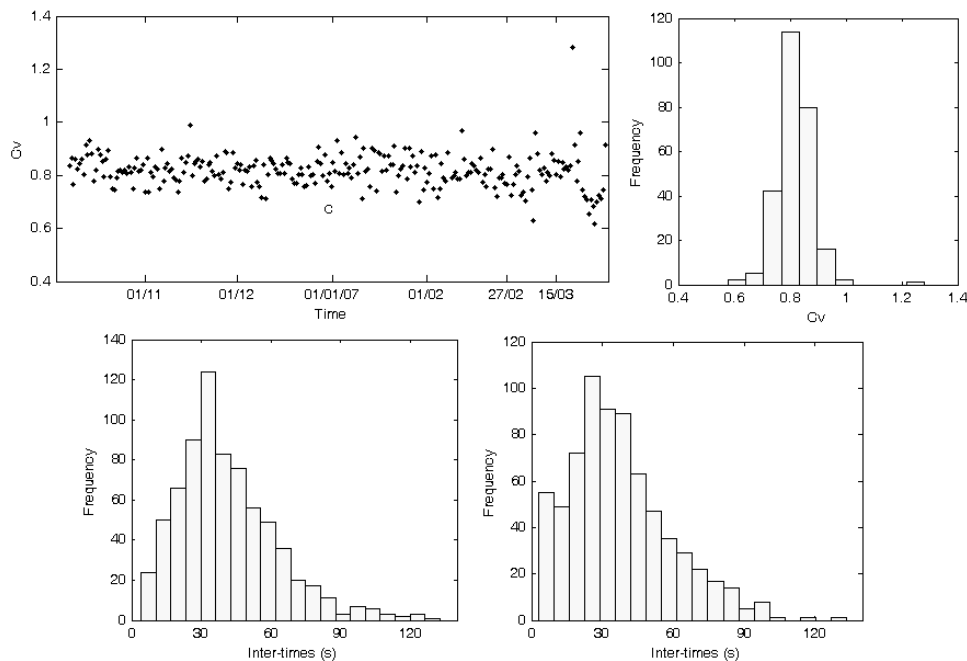


Figure 8.a. Evolution and (b) distribution of the variability coefficient of the inter-time series. In the lower panel: distribution of the inter-times of the swarms of (c) 07/03/2007 and (d) 22/03/2007. Modified after De Martino et al. (2011b).

De Martino et al. (2011 a,b) estimate the energy of each explosion by integrating the squared Hilbert transform of the corresponding seismic events. The limits of the integration are chosen on the basis of the amplitude of the explosion-quakes with respect to the background signal [see De Martino et al., 2004].

Relatively TP1, Figure 6a describes the time-evolution of the mean value and of twice the standard deviation of the energy (in log scale) computed in time-windows of 12 hours. Energy has a complex behaviour with fluctuations on many time scales. Nevertheless, a relatively low-energy can be distinguished until October 2002 (Phase I-II). An high-energy phase occurs in the period October-December 2002, when the mean value increases of about 10 % (Phase III). T-test comparing these two subsets assures that the mean values of the low-energy and high-energy phases are statistically different with a significance level of 99 % or more (De Martino et al., 2011a). In Figure 9 we report energy distributions computed in four different time intervals (De Martino et al., 2011a). Parts (a)-(c) refer to time-intervals before the crisis (Phase I, II, III, whereas Figure 9d is relative to a time-interval during the crisis (Phase IV). As is typical for Strombolian explosions, energy distributions are well described by log-normal distributions [De Martino et al., 2004]. A higher discrepancy between peak value and mean value (μ) and a greater standard deviation (σ) appear in Figure 11c that corresponds to the time-interval during which many spikes of energy occur (Figure 9a). This implies a sharp asymmetry of the distribution, with the values below the peak that are less probable and a deviation from the log-normal distribution.

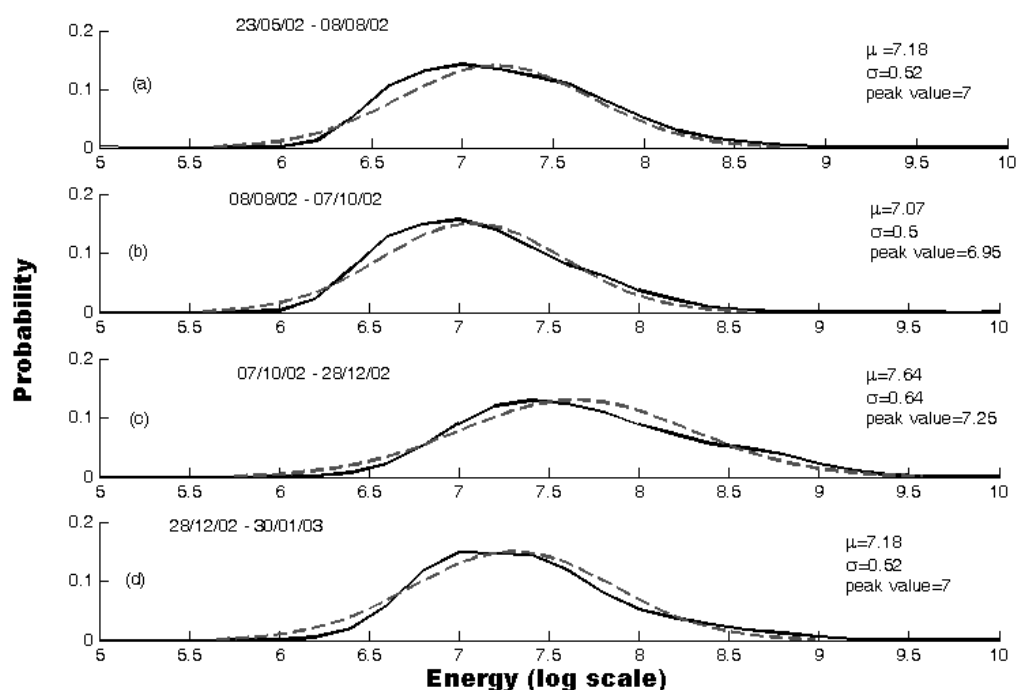


Figure 9. Energy distribution estimated in four disjointed time intervals indicated at the top and corresponding to phases I, II, III, and IV. Phases I, II, and III include three sectors of 60–80 days spanning from the start of recordings to the crisis onset; phase IV includes about 1 month after the crisis onset. Frequency of the solutions is expressed as a probability. Energy is measured in log scale. Modified after De Martino et al. (2011a).

Moreover, in this time-interval the maximum of the distribution moves towards higher values confirming that the energy increases as inferred from its time-evolution. The similarity of the waveforms among the explosions provides an energy distribution with a log-normal shape. This implies a preferred length scale in the source mechanism [De Martino et al., 2004]. According with the cited coalescence Landau-Chandrasekar model [Chandrasekar, 1943], the scale has been identified with the size of the bursting slugs; it depends on specific rheologic parameters and on the rate of inter-times distribution [Bottiglieri et al., 2005]. Looking at 2007, the analysis of the amplitude distributions of the explosion-quakes during NEP shows that at every time they continue to retain a classical log normal behaviour, even when a growth of 35 % is observed at the beginning of January (see Figure 10). The growth in amplitude does not correspond to a simultaneous growth of the occurrence rate, which remains indeed constant, while the Landau-Chandrasekar model expects that longer inter-times provide more energetic explosions. During EP the amplitude distribution shape drastically changes, from a Gaussian to a broader distribution, reflecting the strong variability of the explosions. This is a further indication that the system is out of equilibrium and many scales are reliable.

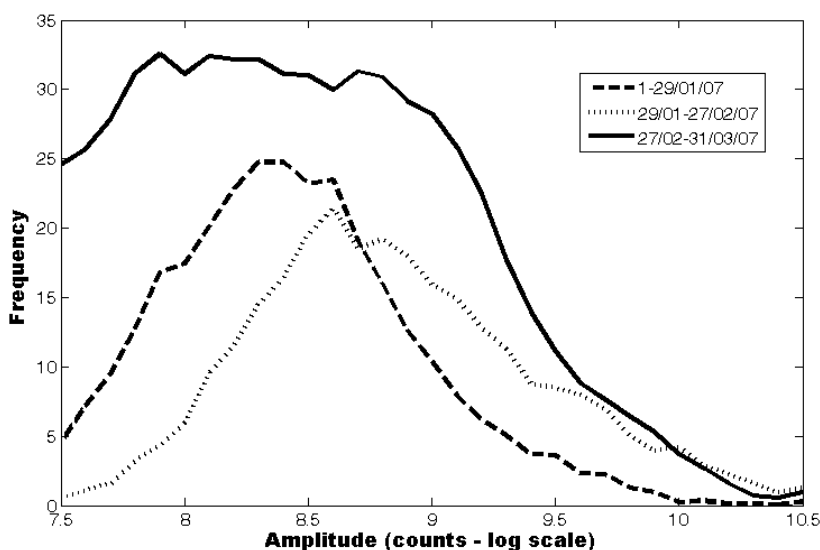


Figure 10. Mean density distributions of the explosions' amplitude estimated within the periods indicated in the legend during TP2. Frequency is the mean number of events occurring in 16 hours. Modified after De Martino et al. (2011b).

EVIDENCE OF A PRECURSOR: A 3-DAY SIGNAL

In the stationary phase, the similarity of the waveforms among the explosions provides an energy distribution with a log-normal shape. This implies a preferred length scale in the source mechanism [De Martino et al., 2004]. According with the cited coalescence Landau-Chandrasekar model [Chandrasekar, 1943], that scale has been identified with the size of the bursting slugs [Bottiglieri et al., 2005] that depends, other than specific rheologic parameters, also on the inter-times.

As is shown in Figure 10, until the onset of the effusion, the analysis of the amplitude distributions of the explosion-quakes shows that at every time they continue to retain a classical log-normal behaviour, even when a slight growth is observed at the beginning of January. The growth in amplitude does not correspond to a simultaneous growth of the occurrence rate, which remains indeed constant, while in stationary equilibrium conditions we expect that longer inter-times would provide more energetic explosions.

During the effusive phase, the amplitude distribution shape drastically changes, from a Gaussian to a more “uniform” behaviour, which reflects the strong variability of the explosions. This is a further indication that the system is out of equilibrium and all the scales are reliable.

The evolution of such phenomenology can be traced by looking at surrogated data, suitable extracted from the data-set. De Martino et al. (2011b) obtained a surrogated series by selecting the highest amplitude among all the explosions occurring in sliding time windows 3 hours long.

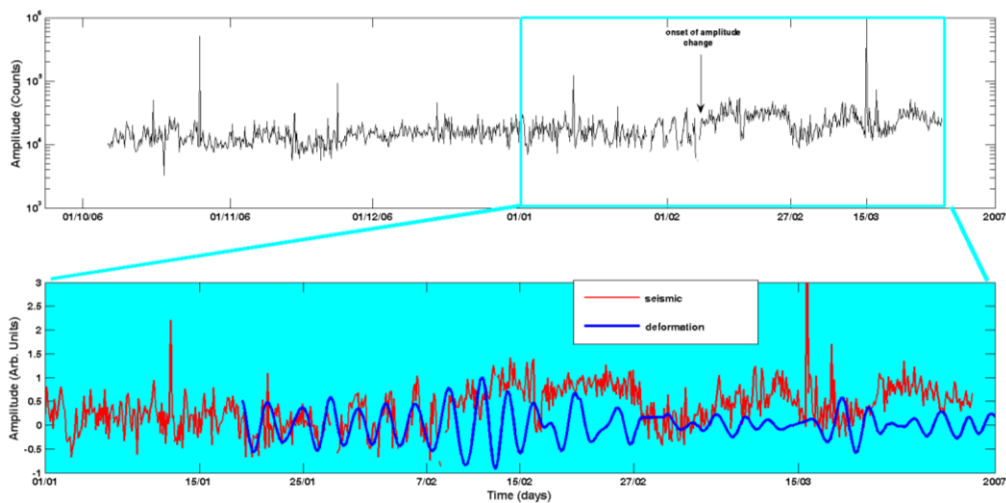


Figure 11.a. Logarithm of the maximum amplitude of the seismic signal estimated in non-overlapping windows of three hours. The highest value represented in the figure corresponds to the big explosion occurred on 15th of March during the effusion; b) Evolution of the amplitude of the seismic signal (red line) along with the signal of the dilatometer filtered between the frequencies corresponding to the periods of two and four days, respectively (blue line). Both the signals have been normalized to facilitate the comparison. Dilatometer's recordings start on January 19th. In this representation, the match between the curves evidenced by a black rectangle lasts approximately from the beginning of the dilatometers' recordings to February 7th, when an increase of the explosions' amplitude occurs and leads to the effusion.

The surrogated data are reported in Figure 11a: the amplitudes' growth is clearly visible as a step function in correspondence of February 5th; the energy involved in the dynamics changes abruptly. This representation clearly shows that this change is introduced by a precursor signal, which appears as a modulation in the explosion amplitudes with a characteristic period of about 3-days. In other words, from 25 of January the amplitudes of the explosions are modulated by a 3-day signal that ends on 7 February. This signal is in the tidal regime and requires to be better investigated by the strainmeter which is more suitable to measure these so low frequency bands. Since the two instruments (seismometer and

strainmeter) measure different physical variables (velocity and strain), the two series were properly normalized (De Martino et al., 2011b).

On the other hand, our interest is only to observe the correspondence of such a 3-day signal on the two instruments. The superposition between the surrogated data from seismic series and the strain signal is reported in Figure 11b. As one can see, the best overlap occurs just in the identified period, highlighted by the box. This correspondence means that a deformation is associated with this precursor. Signals in the tidal frequency range are not usually considered as eventual volcanic precursors, although some relations between tides and seismic vibrations have been studied (see, e.g., Cochran et al., 2004).

To clarify the contribution and the link between the tides and the volcanic activity, De Martino et al. (2011b) analyze the behaviour of the standard tidal constituents together with our precursor in comparison with the salient dates of the non-equilibrium phase.

Figure 12 shows the time evolution of the spectral peaks of earth tides: specifically diurnal, semi-diurnal and 3-day constituents recorded by the strainmeter. Each behaviour is compared with the mean value and the mean value plus 3 (standard deviation), black dash-dot and gray dotted line respectively, both evaluated one year after the end of the effusion (i.e. February-March 2008). Note that the zero corresponds to 01/01/07.

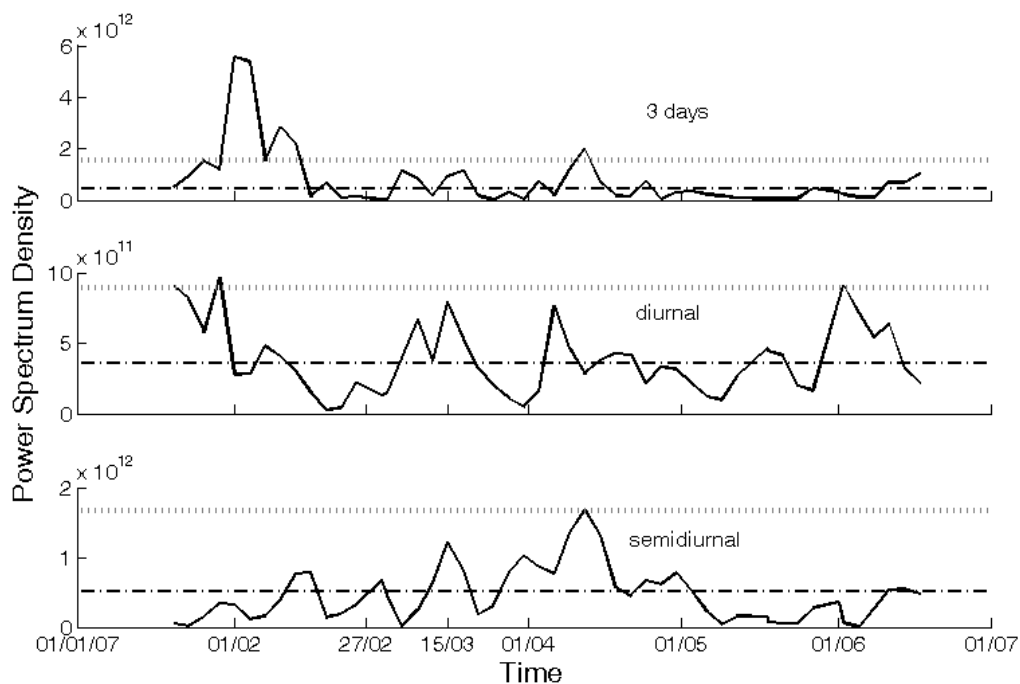


Figure 12. Power Spectral Density (PSD) of the frequencies corresponding to periods of (top) 3 days, (middle) 24 hours and (bottom) 12 hours of the dilatometer in the period January-June 2007 (blue lines). They have been computed by Fourier analysis on windows of 12 days with an overlap of 75 %. All values fall within the statistical variability except the 3-day component in the period January 25th-February 5th 2007, suggesting a significant deviation of the energy of this frequency only in this period. This has been verified by evaluating the mean values (black dash-dot line) and the variability (mean value plus three standard deviation -gray dotted line) of PSD along two months of standard Strombolian activity, namely one year after the end of EP. Modified after De Martino et al. (2011b).

Diurnal and semidiurnal constituents oscillate around the mean value which is basically constant and not connected to the volcanic activity. In addition, both components are always contained in the variability (lower than green line). The evident oscillations are mainly related to the lunar and semilunar tides. On the contrary, the 3-day component shows a clear growth at the end of January 2007 and it returns to the stationary value well before the end of the effusion. This is another indication that it is related to the volcanic activity and it can be considered a transient precursor of the volcanic eruption.

An oscillation with similar frequency content has been observed in the 2002-2003 crisis onset, but the sensor is not suitable for quantitative estimates. Nevertheless, some oscillations in the occurrence rate with a periodicity of 3.5 days in the time interval 15 September 2002 to 5 October 2002 are clearly highlighted in Figure 6c by the rectangle.

CONCLUSION

We have reviewed the seismic wavefield and the statistical and macroscopic features of the Strombolian explosions during the transition from the standard Strombolian activity in the two recent eruptive events.

Regarding TP1, before the crisis, three bell-shaped classes of spectra with maxima falling in the range 1-5 Hz can be recognized (De Martino et al., 2011a). Spectral content has two main changes, the most prominent occurring at the crisis onset when the frequency peak at 0.3 Hz increases in amplitude. Energy of the explosions is lognormally distributed, except during a 2 month time interval before the crisis when it also shows a higher mean value. The inter-occurrence time distributions display a homogeneous Poissonian behavior with a mean intertime of 250 s, without changes at the crisis onset. Only swarms of explosions are not ruled by a Poisson process and display higher occurrence rates and higher energies.

Regarding TP2, see De Martino et al. (2011b), the intertimes follow the same Poissonian distribution, indicating that the degassing layer emits bubbles with the same dynamics, just changing the rate that increases approaching the effusion phase from a mean value of 0.004 to 0.009 (1/s). The increase of the rate (i.e. reduction of the inter-times) is in line with the depressurization due to the effusion. A departure from the Poissonian description of the occurrence of the explosions regards the swarms that occur only during the crisis. In fact, the variability coefficient approaches 0.6 indicating the existence of a most probable inter-time of 30s. Furthermore, the spectra change from a bell shape to a decreasing one having the fundamental frequency at 0.3 Hz when the crisis starts. Energy is lognormally distributed until the crisis onset, with an increase of the mean value when the crisis is approaching becoming flatter after. A precursor 20 days before the crisis is also detected, which is a modulation with a period of 3 days of the amplitude of the explosions, that was studied by analyzing the recordings of both a seismometer and a dilatometer. This signal is followed by an increase of the mean value of the energy which, in turn, introduces the crisis. A similar precursor (an oscillation with similar frequency content) has been observed in the 2002-2003 crisis onset, but the sensor is not suitable for quantitative estimates (De Martino et al., 2011a).

Both the 2002-2003 and 2007 crises can be ascribed to a response of the volcanic system to rising magma visible in the seismic observations. During these crises, approximately the same amount of lava has been emitted [Landi et al., 2009], but the duration of the 2002-2003

crisis is 6 times longer than that of the 2007 eruption. De Martino et al. (2011a) interpret this evidence as the effect of different velocities of magma ascent, inducing low or fast variations of the internal state. It means that in 2007 there is a short transitory phase, whereas in 2002 the transitory phase is long, allowing a smooth response of the system. This could explain the existence of precursors with different timing in the two cases and why in 2007 the effusion, the swarms, and the occurrence rate increase do occur simultaneously, while in 2002–2003 they do not. All these phenomena may be interpreted as the response of the volcano to restore the equilibrium condition perturbed by the injection of new magma. Taking into account analogies and differences between the two crises, the authors depict a possible scenario of the phases involved before and during the crises of Stromboli volcano:

1. A modification of the equilibrium conditions in the plumbing system is induced by rising magma some months before the crisis. This phenomenon is detectable as a modulation of the amplitude (2007) or a frequency change (phase A in 2002–2003) of the explosions.
2. The transitory phase involves an increase of energy as a first soft response of the system. In 2007 it is the unique effect and the lognormal behavior is preserved. In 2002–2003, the energy increase is accompanied by swarms, and lognormality is lost (phase III) suggesting that the departure from the equilibrium has already begun.
3. During the crisis the system is completely out of equilibrium. In 2007 there is a strong response involving simultaneously the effusion, an overall increase of the occurrence rate of the explosions, the occurrence of swarms, and the modification of the usual lognormal distribution of the energy.

In 2002–2003 the response is soft: during the effusion the occurrence rate does not increase and the energy distribution recovers its lognormal shape (phase IV). In both the cases, an overall pressure decrease in the plumbing systems induces a deep gas slug nucleation, with the excitation of low frequencies during the slug ascent (phase C).

The actual departure from that stationarity seems to be traced by a precursor evidenced in the tidal regime (De Martino et al., 2011b). It appears as a transient oscillating signal at about 3-days that modulates the explosion amplitudes. The successive activity can be interpreted as the response of volcano to restore the equilibrium condition. Indeed, variations in rheological characteristics can induce changes in the amplitude ratios among the standard frequency peaks and/or enrichments in the tidal spectrum, both at lower and higher frequencies. In this framework, earth tides observed at Stromboli can be associated with the local rheological characteristics that synchronize with the global aspects. We can think at Stromboli plumbing system as a sort of reservoir that during stationary activity receives and ejects, on average, the same quantity of magma. If the loading rate of that reservoir rapidly increases, the system tends to be choked at a certain depth. The theory of vibrating cavities provides that closed fluid-filled cavities develop lower harmonics than open ones. This should explain in an heuristic way the insurgence of the 3-day signal. The rheological change due to a different magma loading, observed as the 3-day signal in spectrum in the tidal regime, has the consequence to modify the pressure conditions along the volcanic conduit. All the successive activity is the response of volcano to restore the equilibrium condition. New insights on the role of the swarms in the departure from the equilibrium have been recently investigated by De Martino et al. (2012). They interpret these events as a very sensitive proxy for the

escaping from the equilibrium conditions in volcanic systems. Similarly, Ciaramella et al. (2011) have detected the existence of a huge amount of LP events in the Campi Flegrei caldera, whose characteristics revealed by Falanga and Petrosino (2012) and De Lauro et al. (2012) show another effect of local disequilibrium conditions in a different fluid-solid system, i.e. hydrothermal system.

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Chapter 6

VOLCANIC HAZARD ASSESSMENT OF PURACÉ VOLCANO, COLUMBIA, BASED ON STATISTICAL SURVIVAL ANALYSIS OF THE HISTORICAL ERUPTION RECORD: A TUTORIAL

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ABSTRACT

The most reliable volcanic hazard assessment is performed through continuous multi-parameter surveillance of the volcano, recording seismicity, degassing, deformation etc. In many cases, this allows for early alert with regard to impending eruptions. While not developed to substitute for monitoring efforts, the statistical analysis of eruption time series has advanced to become an important component of hazard assessment. It provides a tool for probabilistic eruption forecasting, being the only available approach where no permanent monitoring is feasible. In the ideal case of surveyed volcanoes, it complements the investigation of interacting geophysical, volcanological and geochemical processes. Moreover, it provides an approach to estimate the hazard posed by a volcano on a longer time scale, independently of its present state of activity, and can give an idea if the present observed activity is characteristic for the historically known behaviour of the volcano.

There has been growing awareness for the importance of understanding a volcano's chronological character. However, statistical analyses are not yet systematically performed in the course of hazard assessment, which might in part be due to insufficient instruction on the mathematical handling of the method. The aim is to promote eruption time series analysis in science education, and to establish the statistical techniques such

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that they will be routinely applied to volcanoes by scientists and local authorities who are in charge of volcanic hazard evaluation.

This chapter means to give an introduction into the basics of statistical eruption forecasting, which is based on the analysis of historical eruption time series. The mathematical method is standard knowledge in the realms of medical statistics and engineering applications, where it is known as life-time or failure analysis. In volcanological contexts, however, most scientists are still unfamiliar with this approach; in particular, the Kaplan-Meier estimate has only recently been proposed as an improvement of the method (Dzierma and Wehrmann, 2012). The basic ideas are presented in the form of a tutorial applied to the historical eruption record of Volcán Puracé, Columbia, to provide a readily understandable guideline on how to perform the individual steps of the procedure and interpret the results of the statistical analysis. Puracé belongs to the most active volcanoes of the Columbian volcanic chain, which is generated by subduction of the Nazca Plate beneath the South American Plate. Its recent eruptive history is of mainly explosive behaviour, and 23 eruptions have been recorded since 1820.

INTRODUCTION

The most efficient and reliable volcanic hazard assessment is performed through continuous multi-parameter surveillance of a volcano, recording seismicity, degassing, deformation, etc. (e.g., Scarpa and Tilling, 1996). Changing volcano-seismic signals can give insight into the movement of magmas and gasses within the volcanic edifice, and can be representative of the present state of activity of a volcano (Wassermann, 2002; Chouet, 2003; McNutt, 2005, and references therein). Both the form of the seismic signals (high-frequency vs. low-frequency events, and tremors) and the amount of activity measured by the real-time seismic amplitude are indicative of changing processes and increasing state of unrest (e.g., Endo and Murray, 1991; Chouet, 1996a,b; McNutt 1996, 2000a,b, 2002). Magma ascent and intrusion may furthermore be reflected by inflation and deformation of the volcano recorded by techniques ranging from simple “naked eye” observations (in the more sophisticated case using real-time video monitoring) over tiltmeters and extensometers to radar interferometry measurements by satellite (e.g., Linde et al., 1993; Dvorak and Dzurisin, 1997; Voight et al., 1998; Murray et al., 2000; Dzurisin, 2003; Pritchard and Simons, 2002, 2004; Masterlark and Lu, 2004). The additional mass of intruded magma may be detected in a microgravity study (e.g., Rymer, 1996; Gottsmann and Rymer, 2002; Williams-Jones and Rymer, 2002). From the geochemical point of view, frequent sampling of hydrothermal fluids and gas emissions from the volcano – and the crater lake, where it exists – can give insight into changes in temperature, acidity and chemical composition which may help to decide whether a degassing volcano is fed by fresh magma at depth or just showing increased activity due to, for instance, rain water entering the crater and released as steam (Martini, 1996; Stix and Gaonac’h, 2000; Zimmer and Erzinger, 2003). The geochemical composition of degassing clouds may also be remotely investigated based on the composition-specific spectral absorption at different frequencies – either with ground-based stations or from space (e.g., Khokhar et al., 2005; Stix et al., 2008; Galle et al., 2010). By combining a number of these different approaches, over many years monitoring experience a good picture of the characteristic volcano properties and behaviour will emerge and can in many cases allow for early alert with regard to impending eruptions (Swanson et al., 1983; Malone et al., 1983; Dzurisin et al., 1983; Pinatubo Volcano

Observatory Team, 1991; IAVCEI Subcommittee on Decade Volcanoes, 1994; also see Tilling, 2008 and references therein).

There are two major set-backs with this approach to volcano hazard mitigation. The first is of a purely practical nature: the man-power and cost of maintaining such a detailed monitoring network for all potentially active volcanoes in the world is far beyond what is realistically feasible. The instrument cost for implementing such an extensive monitoring campaign is enormous, and so is the number of people necessary to install and maintain all the monitoring devices. Many volcanoes are all but inaccessible, and even where the planned installation site can be reached easily by car or horse (or helicopter), they are usually remote from civilization and the electrical supply for the instruments will pose a problem – not to mention the real-time telemetry of the measurement results to the responsible volcano observatory. Inhospitable conditions will necessitate frequent repair and maintenance visits to keep all stations functional in the face of weather, animals, and the volcano activity itself (e.g. volcanic bombs, rock-slides or acid gases). Particularly in countries with a large number of volcanoes, only the most active – and most populated – can possibly be monitored.

The second issue is perhaps the most severe, because it would apply even in an ideal world in which constant, real-time multi-parameter monitoring were implemented for each and every volcano, no matter where. It is the time-scale applied in hazard assessment. Real-time monitoring is an essential tool when it comes to timely recognition of increasing risk, and can allow for early warning of surrounding populations. The decision to evacuate can only be based on monitoring and past experience. What monitoring cannot do, however, is estimate the volcano hazard over long time scales in the past and future, even when it is presently in a quiet state. Monitoring will paint an instantaneous picture of the volcano behaviour at present, but without knowing if this is in any way representative of the eruptive pattern over time scales exceeding several decades. In particular, it cannot predict the likelihood of future eruptions until they are immediately imminent. When it comes to decisions of the kind, whether a tourist resort or a major roadway can be established on a volcano flank, the probability of eruptions many years in the future must be considered. To achieve this, statistical analysis of eruption frequency and magnitude is the approach of choice, becoming particularly valuable in combination with hazard maps based on the terrain properties and past eruptions.

This should emphasize that statistical eruption analysis is not meant to substitute for monitoring efforts, but to provide a tool for probabilistic eruption forecasting. In cases where no permanent monitoring is feasible, it is the only available approach for hazard assessment. But even in the much preferable case of instrumentally monitored volcanoes, it can complement the monitoring with a statistical hazard estimate on a longer time scale, independently of the present state of unrest. This approach can help assess if the presently observed activity is characteristic for the historical (or even geological) record of volcano activity.

In recent years there has been a growing awareness of the importance of understanding a volcano's chronological eruption behaviour through statistical means. After the pioneering work of Wickman (1976) and Klein (1982), statistical analyses have been performed for a variety of volcanoes (Ho, 1990,1991; Bebbington and Lai, 1996a, b; Connor et al., 2003; Bebbington, 2008; De la Cruz-Reyna, 1991,1996; Mendoza-Rosas and De la Cruz-Reyna, 2009,2010; Dzierma and Wehrmann, 2010a,b; Sobradelo et al., 2011). However, statistical analyses are not yet systematically performed in the course of hazard assessment, which

might in part be due to insufficient instruction on the mathematical implementation of the method. The aim of this chapter is to promote eruption time series analysis, and to describe the appropriate statistical techniques such that they can be routinely applied to volcanoes by scientists and local authorities who are in charge of volcanic hazard evaluation.

This chapter provides an introduction to the basics of statistical eruption forecasting, which is based on the analysis of historical eruption time series. The mathematical method is standard knowledge in the realms of medical statistics and engineering applications, where it is known as life-time or failure analysis. In volcanological contexts, however, most scientists are unfamiliar with this approach; in particular, the Kaplan-Meier estimate has only recently been proposed as an improvement of the method (Dzierma and Wehrmann, 2012). The basic ideas will be presented in the form of a tutorial applied to the historical eruption record of Volcán Puracé, Columbia, to provide a readily understandable guide for performing the individual steps of the procedure.

GEOTECTONIC AND VOLCANOLOGICAL SETTING

The Colombian volcanoes form part of the Andean Cordillera, an over 7500 km long morphologically continuous mountain chain formed since the early Jurassic on a Proterozoic to Early Mesozoic Basement (Stern, 2004). All along the western margin of South America, the Andean mountain-building and active volcanism are maintained by subduction of the Nazca Plate (Colombia to South Chile) and Antarctic Plate (south of the Chile Triple Junction at 46.5°S) beneath the South American Plate, with convergence velocities of 7-9 cm/yr for the Nazca Plate and 2 cm/yr for the Antarctic Plate, respectively (DeMets et al., 2010).

Active volcanism occurs along four segments of the Andean arc: the Northern Volcanic Zone in Colombia and Ecuador (NVZ, 5°N-2°S), the Central Volcanic Zone in Perú and Chile (14-27°S), and the Southern Volcanic Zone (33-46°S) and Austral Volcanic Zone in Chile and Argentina (49-55°S) (Stern, 2004; for a recent overview of Andean volcanic hazards, compare Tilling, 2009). The interruption of active volcanism between these zones has been attributed to the subduction of ridges giving rise to flat-slab segments with near-horizontal subduction below 100 km depth (Gutscher et al., 2000; Espurt et al., 2008).

The Northern Volcanic Zone includes 19 volcanoes in Colombia (Méndez Fajury, 1989) and at least 20 volcanoes with Holocene activity in Ecuador (Hall et al., 2008), aligned approximately north-south. Among the Colombian volcanoes, the best-known are Nevado del Ruiz, with its devastating lahar in 1985 with ca. 23 000 fatalities (Naranjo et al., 1986; Mileti et al., 1991), and Galeras volcano, one of the most active volcanoes in Colombia (Fischer et al., 1995; Baxter et al., 1997; Cardona, 1997). Puracé volcano, though also one of the most active in Colombia, has received less attention, possibly because no eruptions with fatalities have occurred since 1949. The last known eruption was a central vent explosive eruption in March 1977 (Global Volcanism Program, GVP, Simkin and Siebert, 1994; Siebert and Simkin, 2002).

Puracé is the northernmost volcano of the Coconucos Volcanic Chain, comprising 15 volcanic centres oriented N40°W (<http://www.ingominas.gov.co/Popayan/Volcanes/Volcan-Purace/Generalidades.aspx>; Oppenheim, 1950; Arcila, 1996). Puracé is situated at 2°19'01 N, 76°23'53 W in the Cauca department (Figure 1), and is 26 km from the town of Popayán,

with over 200,000 inhabitants. Two rivers descend from its flanks, Río Vinagre and Río San Francisco, the deeply cut valleys of which provide paths for pyroclastic flows or lahars. The volcanic edifice is a truncated stratocone 4650 m high, mainly made up of andesitic lava flows interbedded with andesitic tuffs, ash, bombs and blocks, built on top of a dacite shield volcano. It has concentric craters 500 and 900 m in diameter with fumarolic activity both inside the crater and on the NW flank of the volcano.

Puracé has produced a number of mainly explosive eruptions with pyroclastic flows and tephra fall-out, making it one of the historically most active volcanoes of Colombia. Seismic monitoring by the Instituto Colombiano de Geología y Minería (INGEOMINAS) was started in 1986, and the volcano is presently monitored with 6 seismic stations, including a broad-band seismometer (<http://www.ingegominas.gov.co/Observatorios-Vulcanologicos.aspx>). Three tilt-meters are monitoring deformation and ground-motion of the volcano. Fumaroles and hydrothermal springs are sampled and chemically analysed by INGEOMINAS (Arcila, 1996; Serna et al., 2003). The volcano is presently in a quiet state (classified as “green” by INGEOMINAS).

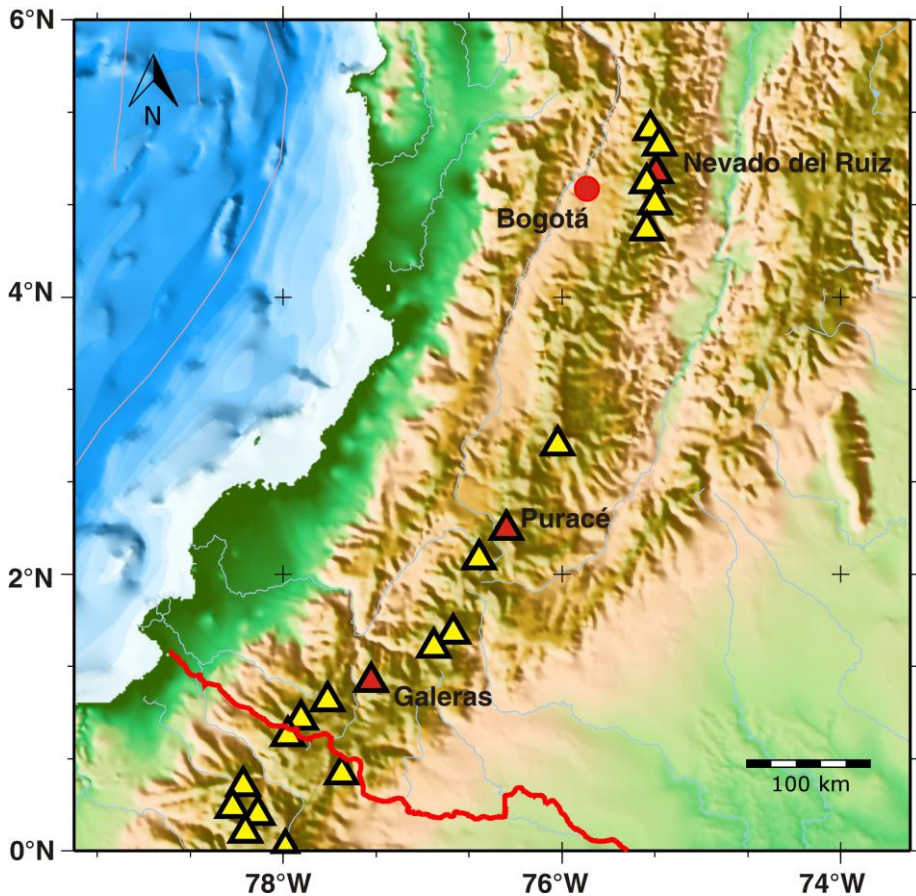


Figure 1. Location map of Colombian Volcanoes (yellow and red triangles). Red line: border to Ecuador.

Table 1. Historical eruption record of Puracé volcano, from the Global Volcanism Program (GVP, Siebert and Simkin, 2002-, www.volcano.si.edu)

Start	Stop	VEI	Eruptive characteristics	Repose time (years)
1816	?	?	?	not used
1827 Nov 18	?	2	central vent eruption, explosive eruption, lava flow(s)	
1835 Jan 23	?	2	central vent eruption, explosive eruption	8
1840	?	2	central vent eruption, explosive eruption	5
1847 Oct 27	1852	3 ?	central vent eruption, explosive eruption, damage (land, property, etc.), evacuation	7
1860 $\pm$ 9 yrs	?	2 ?	central vent eruption, explosive eruption	13
1869 Oct 4	1869 Nov	3 ?	central vent eruption, explosive eruption, pyroclastic flow(s), mudflow(s) (lahars)	9
1870 Oct	?	2	central vent eruption, explosive eruption	1
1878 Aug 31	?	2	central vent eruption, explosive eruption	8
1881	?	2	central vent eruption, explosive eruption	3
1885 May 25	?	3	central vent eruption, explosive eruption, fatalities, damage (land property, etc.)	4
1899	?	2	central vent eruption, explosive eruption	14
1902 (?)	?	2	central vent eruption, explosive eruption	3
1906	?	2	central vent eruption, explosive eruption	4
1924	?	2	central vent eruption, explosive eruption	18
1925 Oct 12	1925 Nov 5	2	central vent eruption, explosive eruption	1
1926 Aug	1926 Sep	2	central vent eruption, explosive eruption	1
1927	?	2	central vent eruption, explosive eruption	1
1946 Mar	1946 Apr	2	central vent eruption, explosive eruption	19
1947 Apr 27	?	2	central vent eruption, explosive eruption	1
1949 May 26	1949 Jun 11	2	central vent eruption, explosive eruption, fatalities	2
1956 (in or before)	?	2	central vent eruption, explosive eruption	7
1957	?	2 ?	?	1
1977 Mar 19	1977 Mar 28 (?)	2	central vent eruption, explosive eruption	20
			no eruption 2011	34 - censored

PURACE ERUPTION RECORD

A statistical eruption analysis is usually based on the historical eruption record of a volcano (in some cases the geological record can also be included, compare Wehrmann and Dzierma, 2011). In the humanities, history is distinguished from archaeology by the fact that written records exist for events, rather than material remnants – this same distinction can be made between the historical vs. geological eruption record. The latter is based mainly on pyroclastic deposits, where explosive eruptions are concerned, of which the chronological sequence is constrained from stratigraphic relationships, radiometric dating, etc. Aside from this difference in tradition and dating method, in many cases historical eruptions are classified as those which occurred in “the recent past” (in which historical or oral records might have

existed), even if a given eruption is dated on the basis of tephrostratigraphy or by other means.

A good record of eruptions is the basis for a meaningful statistical analysis, since any missing or mis-classified eruptions will influence the outcome of the statistical method. The analysis presented here is based on data from the published literature only.

The Smithsonian Institution Global Volcanism Program (GVP) offers historical and geological eruption records for virtually all known active volcanoes (Siebert and Simkin, 2002), and although uncertainties cannot be excluded, this is a valuable starting point in the search for volcano eruption histories and other information. For Puracé, the statistical method will be presented based on the record in the GVP database, given in Table 1. The first known historical eruption occurred in 1816 AD.

No further details such as stop date, eruptive characteristics, or VEI are given, so it will not be included in the analysis.

The volcano explosivity index (VEI) (defined by Newhall and Self (1982) and refined by Simkin and Siebert, 1994) is a measure of the eruption magnitude for dominantly explosive eruptions. It relies mainly on the erupted tephra volumes, but includes several other parameters such as eruption duration, column height, and qualitative descriptions. For statistical analyses, it is important that only eruptions of similar characteristics are included; this is mostly done by excluding eruptions below a particular VEI threshold. The smaller eruptions are at a higher risk of being missed, so the eruption record may be incomplete and biased toward the larger events.

Very large eruptions, however, are very rare; therefore, a compromise must be made. For most practical purposes, a VEI threshold of $VEI \geq 2$ is a reasonable choice – if the record permits, $VEI \geq 3$ will give a more reliable data set in the sense that $VEI \geq 3$ eruptions are less likely to be incomplete (missing in the record) than the smaller $VEI \geq 2$ eruptions (e.g., De la Cruz-Reyna, 1991), but in cases such as Puracé, this is impossible: with just three eruptions of possibly $VEI = 3$, no statistical analysis would be feasible. It might be worth investigating whether other measures of eruption strength (in particular, magnitude) can be useful for the choice of eruptions to include in the data set. I here adhere to the general method used in statistical volcanology, which bases the data selection on VEI.

For many eruptions in the historical record, no stop date is registered; in the case of a slowly waning eruption, it is hard to put a definite end date to eruptive activity. In the statistical analysis, the stopping date is not necessarily a vital problem; since eruptions are treated as “point events”, the eruption duration is neglected in the analysis and the eruption date is taken as the onset of each eruption. Accordingly, the repose time – i.e., the inter-eruption time – is defined as the time between the onsets of two successive eruptions (Klein, 1982). However, it must be kept in mind that this mathematical simplification carries some problems when eruptions are of long duration or even uncertain end.

For unknown stopping dates, eruptions occurring closely after each pose a difficulty in deciding whether to consider them as one eruption with renewed strength, or as two separate events. Even more severely, ongoing eruptive activity over years or decades make any attempt of eruption likelihood prediction invalid if it is based on only the onset time of activity and treating the volcano as though it had been quiet ever since. In the case of Puracé, all but one eruption with known termination were of short duration of the order of days to weeks (the only exception is the 1847 VEI 3? eruption), which can be neglected in comparison with the time between eruptions (one year or more).

The statistical analysis for Puracé will hence be based on the historical eruption record from 1827 until today (2011), the majority being central vent explosive eruptions of short duration, with the last one occurring in 1977.

The following paragraphs will present the main steps of the analysis. After some basic tests which must be made to check that the eruption record is suitable for this kind of statistical analysis, the volcano repose times are used to calculate the survival function by means of the Kaplan-Meier estimator. This is then fit by simple parametric distributions (exponential, Weibull and log-logistic).

The goodness of the fits must then be ascertained, and the best model of the three is selected based on information criteria. Having obtained a good model for the survival function, the probability of future eruptions can be calculated from the residual life distribution function.

BASIC PREREQUISITES FOR STATISTICAL ERUPTION ANALYSIS

Before embarking on the statistical eruption analysis, several preliminary checks need to be performed to assess whether the method is applicable to the volcano in question. In particular, the quality of the data set must be tested and it must be ascertained that the volcano behaviour is sufficiently uniform so that the past eruption record is complete and in some way representative for the future behaviour.

As a first step, we check for stationarity of the time series, which means that the probability distribution or parameters describing the eruption time series do not depend on the time interval considered. If the volcano behaviour is not stationary, there is no reason to believe that the future (and present) behaviour of the volcano will resemble the past eruption behaviour; in this case, there would be no use in fitting the past repose times with the aim of predicting future eruption likelihoods.

In a case where the volcano eruptions do not follow stationary behaviour, advanced techniques can (and must) be applied which rely on identifying different phases of enhanced/reduced volcano activity and taking into account the switching between these regimes in the analysis (e.g., Bebbington, 2007; Mendoza-Rosas and De la Cruz-Reyna, 2008, 2009).

A first visual impression of the volcano eruption record can be gained by plotting the cumulative number of eruptions vs. time (Figure 2). An ideal, steadily and regularly erupting volcano would result in a step-shaped pattern close to a straight line with slope “average number of eruptions per year” – equivalent to “total number of eruptions over the duration of the historical record”.

For Puracé, this is the case for eruptions until 1977 (red line) – after this, no eruptions are known to have occurred, resulting in a seemingly unusual quiet phase and some deviation from this ideal regular behaviour (approximated by the green line).

Apart from this recent inactivity, the eruptions appear to have followed a rather regular pattern over almost 200 years, a good omen for the statistical analysis. In particular, there is no break in the slope at early times in the eruption record, which sometimes points to incompleteness at early times in the record.

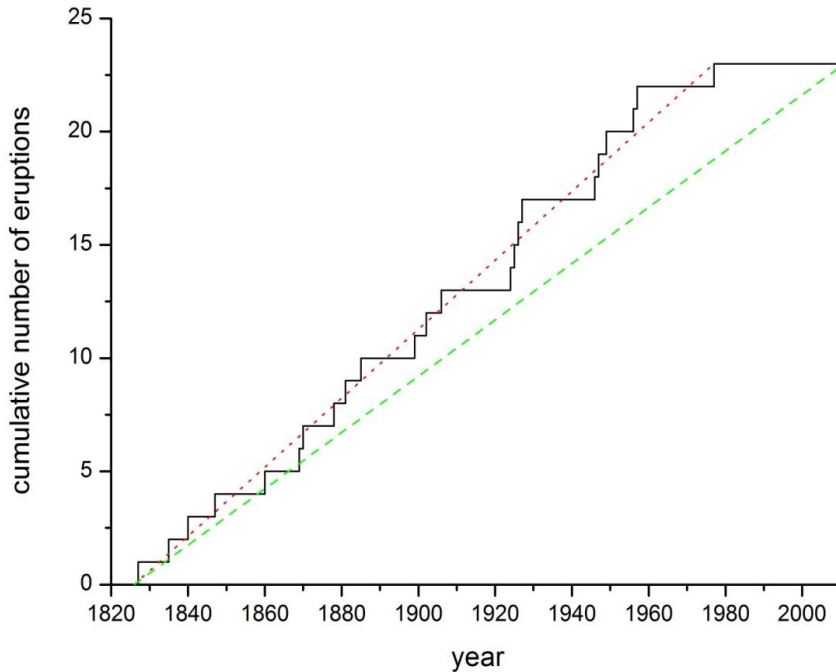


Figure 2. Cumulative number of eruptions (VEI ≥ 2) of Puracé volcano vs. time. The green dashed line shows a hypothetical case of constant eruption rate from the beginning of the record until today (2011); the red dotted line would correspond to a constant eruption rate until the last eruption in 1977.

To confirm the visual impression of a largely stationary volcano behaviour, a moving average test is carried out (Figure 3; Klein, 1982; De la Cruz-Reyna, 1996; Mendoza-Rosas and De la Cruz-Reyna, 2008; Dzierma and Wehrmann, 2010a,b). This concept is similar to control charts applied in industry, where the performance of some system is monitored over time and deviations from “normal functioning” are observed as excursions of the measurements from a given mean value (e.g., Wheeler, 1990, 2000). In volcano statistics, this test is performed by considering the 5-point moving average of the repose times as a function of time. Taking the average over five successive repose times is meant to “average out” short-term statistical fluctuations – for stationary behaviour, it is expected that all the 5-point moving averages should occur close to their mean value (red line). The amount of “normal” statistical fluctuation from one repose time to the next is assessed by the 5-point moving standard deviation – if a systematic change in eruptive behaviour occurs, this should cause the new 5-point average to fall outside the average standard deviation from the mean. Actually, it can be argued that not the total average of all 5-point running averages and 5-point running standard deviations should be taken, but only of all those up to the point in question (Figure 3b). In the case of the Puracé eruption record, for both methods the 5-point averages lie within one standard deviation from the mean, with the exception of the last repose time average, which includes the quiet phase since 1977. A deviation of just more than one standard deviation from the mean can be due to purely statistical fluctuations, but may indicate also in this case that the last long repose time is unusual. Apart from this point, all other repose time averages are randomly distributed about the mean, so the volcano behaviour can be reasonably assumed to have been largely uniform over the historical past.

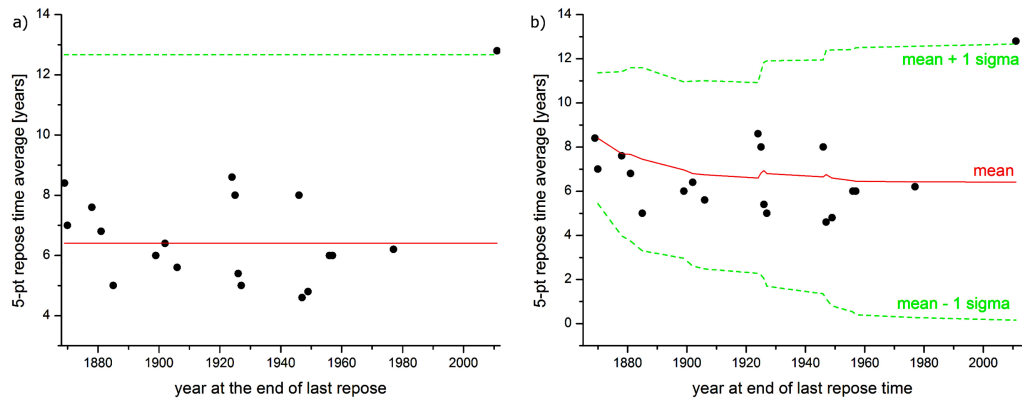


Figure 3. Moving-average test for stationarity. a) In the standard form of this test, the moving average of each five consecutive repose times is plotted against time. The red line is the average of the individual 5-point average values; the green line is one standard deviation from the mean. In this case, the standard deviation is calculated as the average of the individual 5-point standard deviations. b) Instead of taking the average of all 5-point averages and 5-point standard deviations, only the average until the year at the end of the last repose time is taken.

If the eruption time series is approximated by a statistical process, with repose times of random duration, then successive repose times should be independent of each other. This is checked by plotting each repose time as a function of the previous repose and calculating the correlation coefficient (Cox and Lewis, 1966; Mendoza-Rosas and De la Cruz-Reyna, 2008). If successive repose times are uncorrelated, we will assume that they are independent. Whether or not the last repose time is included in the analysis, no significant correlation is found (Figure 4a,b).

From these analyses, the eruption record is deemed suitable for statistical analysis in that successive repose times are uncorrelated, and the behaviour is largely stationary. The last long repose time from 1977 until today is the only marked deviation of the volcano from the usual eruption pattern. In the following, the analysis will be performed from two points of view: today's, including the long last repose time, and secondly if the analysis had been performed in the year 1978.

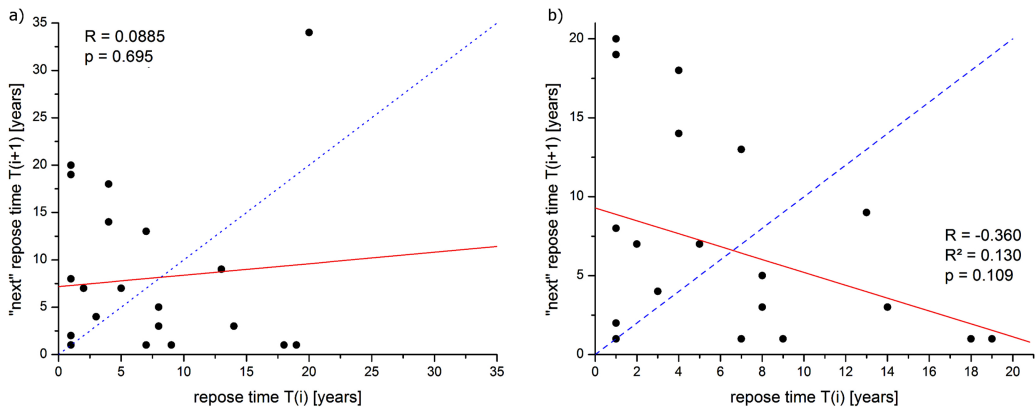


Figure 4. Independence test a) for all repose times, b) for all repose times except for the last (exceptionally long) repose time. In both cases, no statistically significant correlation is observed.

VOLCANO SURVIVAL FUNCTION

For the statistical analysis, we first estimate the volcano survival function using the Kaplan-Meier estimator computed after the observed repose time intervals. Once we have this function, we need to find a statistical distribution which best fits it, which we can then use later to make forecasts. The statistical distributions we consider are the exponential, Weibull and log-logistic distributions. All three are fit to the data using the method of maximum likelihood. We then use the Anderson-Darling and Kolmogorov-Smirnov-tests to estimate the goodness of the fit and the information criteria AIC and BIC to select the best model. This model is then used for the forecast of the statistical eruption probability. Each of the analysis steps will now be considered in some more detail.

For stationary time series, the (cumulative) repose time distribution function $F(t)$ is defined as the probability that the observed repose time T is smaller than or equal to t ,

$$F(t) = P\{T \leq t\} \in [0,1]$$

(e.g., Marshall and Olkin, 2007). The complementary function is called the survival function $S(t)$,

$$S(t) = P\{T > t\} = 1 - F(t).$$

If $F(t)$ is differentiable, the instantaneous eruption probability is given by the density function $f(t)$, with

$$f(t) = \frac{d}{dt} F(t).$$

The survival function can be easily computed from the distribution of repose times as

$$\hat{S}(t) = \frac{\text{number of repose times} \geq t}{\text{number of repose times in the data set}}.$$

If it is calculated in this way, it is known as the “empirical survival function” (e.g., Collett, 2003). The drawback of using the empirical survival function is that it cannot cope with censored data, i.e., with repose times that are not concluded by an eruption. This is the case for the last repose time – after 1977, we know that no eruption has occurred yet, so the repose time must be at least 34 years, but possibly much longer. If the last repose time is discarded, important and unique information is lost for the analysis. The preferable way is to choose a generalized form of survival function which accommodates for censored data: the Kaplan-Meier (or product-limit) estimate of the survival function (Kaplan and Meier, 1958). The calculation of the Kaplan-Meier estimate is shown in Table 2.

For the Kaplan-Meier estimate of the survival function, the intervals are defined by the uncensored repose times. For each interval $j=1\dots k$, the probability of a volcano erupting within this interval is given by the number of repose times within this interval e_j (“events” in Table 2), divided by the remaining repose times n_j (“number at risk”) at the start of the interval. The survival probability through the interval is therefore

$$S(t) = \prod_{j=1}^k \left(1 - \frac{e_j}{n_j}\right) = \prod_{j=1}^k \frac{n_j - e_j}{n_j}.$$

If there are no censored repose times, n_j is the number of repose times remaining in the previous interval, minus the number of eruptions ($n_j = n_{j-1} - e_{j-1}$). In the case of censored data, the censored times c_j are excluded from the next interval ($n_j = n_{j-1} - e_{j-1} - c_{j-1}$). Where no censored data exist, the Kaplan-Meier estimate is identical to the empirical survival function.

Although the Kaplan-Meier estimate of the survival function is relatively new in the context of statistical volcanology (Dzierma and Wehrmann, 2012), it is part of the standard repertoire of medical statistics, and as such routinely calculated by many statistical software packages (among them, e.g., Origin and R) – for a more detailed introduction to the Kaplan-Meier estimate and the calculation of its standard errors (Greenwood’s formula), see Collett (2003) or any other text on survival data analysis. For the purpose of predicting the probability of future volcanic eruptions, it is useful to fit the survival function with a parametric distribution function, i.e. a statistical model described by a small number of parameters. While restricting the range of applicability of the model, this will make predictions more precise, which is what we are most interested in for the volcanological application. Furthermore, the choice of the best model can give us some mathematical insight into the volcano behaviour. The three most generally used models for the survival functions are the exponential, Weibull, and log-logistic distributions:

$$\begin{aligned} S_{\text{exp}}(t) &= \exp\left(-\frac{t}{\tau}\right), \\ S_{WB}(t) &= \exp\left(-\left(\frac{t}{\tau}\right)^\alpha\right), \\ S_{\text{log}}(t) &= \frac{1}{1 + \left(\frac{t}{\tau}\right)^\alpha}. \end{aligned}$$

Table 2. Kaplan-Meier estimate of the survival function

Time (years)	Number at risk	Number of events	Survival	Std. err.
1	23	6	0.7391	0.0916
2	17	1	0.6957	0.0959
3	16	2	0.6087	0.1018
4	14	2	0.5217	0.1042
5	12	1	0.4783	0.1042
7	11	2	0.3913	0.1018
8	9	2	0.3043	0.0959
9	7	1	0.2609	0.0916
13	6	1	0.2174	0.0860
14	5	1	0.1739	0.0790
18	4	1	0.1304	0.0702
19	3	1	0.0870	0.0588
20	2	1	0.0435	0.0425

The exponential distribution corresponds to the simplest case of an ideal Poisson process (e.g., Cox and Lewis, 1966), in which eruptions occur at random, at a constant rate. In this kind of process, events occur

1. seldom (in particular, the probability that two events coincide is negligible),
2. independently of each other (the probability that an event in any time interval occurs is not influenced by any other occurrences before the start of the time interval)
3. with stationary probability (without time trend).

If waxing/waning activity needs to be included in the model, the Weibull distribution provides a model of “simple failure” in which the eruption (= failure) probability is influenced by one dominant process (Ho, 1991; Bebbington and Lai, 1996a,b; Watt et al., 2007). The log-logistic distribution, finally, can represent the interaction of two competing processes (Connor et al., 2003).

FITTING THE SURVIVAL FUNCTION

The models are fit to the survival data (Figure 5) using the method of maximum likelihood to obtain the best parameter estimates – fit parameters are given in Table 3. The goodness-of-fit is estimated by the Kolmogorov-Smirnov test (e.g., Gibbons, 1976) and the Anderson-Darling test (Anderson and Darling, 1952), which are passed by all fits (a large “p” value indicates that there is a large probability that deviations of the fit from the data have arisen by chance).

The choice between the three models can be made on the basis of the Akaike information criterion *AIC* or Bayes information criterion *BIC*, defined by

$$\begin{aligned} AIC &= 2k - 2 \ln(L), \\ BIC &= k \ln(n) - 2 \ln(L), \end{aligned}$$

where k is the number of model parameters, n the number of data points fit by the model, and $\ln(L)$ the log-likelihood function of the estimated model (Akaike, 1973; Schwartz, 1978).

Both the *AIC* and *BIC* evaluate the model by weighting against each other the achieved fit (or remaining misfit) to the data, and the number of model parameters k needed to achieve this fit. The difference between the *AIC* and *BIC* depends on how a large number of model parameters k is penalized. While it is a matter of opinion which information criterion is preferred (compare, e.g., Burnham and Anderson, 1998), both are a useful measure for the relative quality of the models, since they penalize both the misfit and a large number of parameters. The absolute value of the *AIC* (*BIC*) is meaningless, but the model with the lowest *AIC* (or *BIC*) value is the one which achieves the best fit with the smallest number of necessary parameters.

For the Puracé eruption record (both until 2011 and 1978), the *AIC* and *BIC* for the three models is very similar, with only a slight preference for the exponential model. Even without consulting the information criteria, it can be seen from the fit parameters that the Weibull model can be discarded in the following, since its shape parameter is not significantly

different from zero – i.e., it does not differ significantly from an exponential distribution. The log-logistic distribution passes the goodness-of-fit test and has only slightly worse *AIC* (*BIC*) than the exponential distribution. In the following, we will therefore consider the predictions from both models.

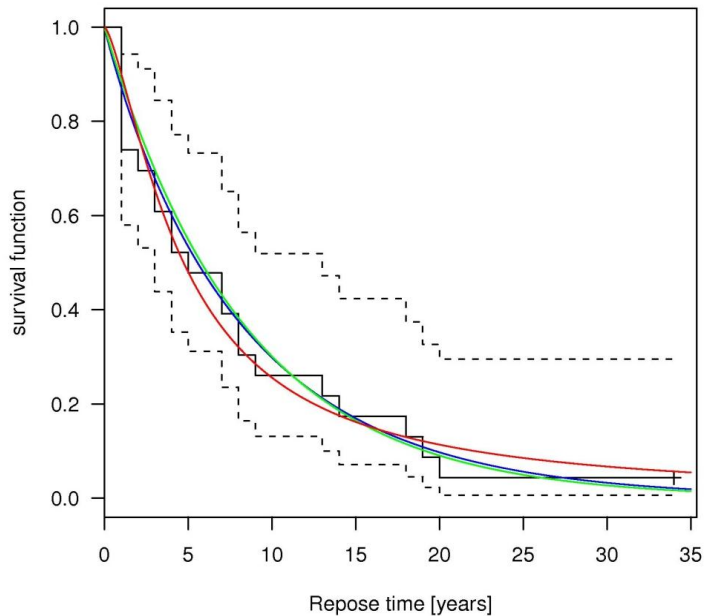


Figure 5. Kaplan-Meier estimate of the survival function (black line, with standard deviation dashed), together with survival distribution fits (green line: exponential; blue line: Weibull, red line: log-logistic distribution).

Table 3. Fit parameters for Puracé survival functions. The fits are performed in R, and fit parameters are related to the definitions used in the text by $\tau = \exp(\text{intercept})$, $a = \exp(-\log(\text{scale}))$

	Intercept parameter			Log(scale) parameter			Loglike- lihood	AIC	BIC	KS- test	AD- test
	Value	Std. err.	p	Value	Std. err.	p				p	p
analysis as performed in 2011											
exponential	2.120	0.219	2.24e-23	0	0	-	-68.7	139.5	140.6	0.697	0.714
Weibull	2.102	0.235	3.37e-19	0.056	0.167	7.37e-1	-68.7	141.3	143.6	0.812	0.783
loglogistic	1.552	0.260	2.46e-09	-0.352	0.173	4.17e-2	-68.1	140.3	142.6	0.584	0.657
analysis, if it had been performed in 1978											
exponential	1.930	0.213	1.64e-19	0	0	-	-64.4	130.8	131.9	0.525	0.591
Weibull	1.962	0.204	7.38e-22	-0.095	0.166	5.70e-1	-64.2	132.4	134.7	0.345	0.430
loglogistic	1.474	0.245	1.77e-09	-0.441	0.171	9.77e-3	-65.2	134.4	136.7	0.252	0.369
“large“ eruptions or clusters											
analysis performed in 2011											
exponential	3.31	0.41	5.35e-16	0	0	-	-25.8	53.7	53.6	0.203	0.48
analysis performed in 1978											
exponential	3.08	0.41	4.26e-14	0	0	-	-24.5	51.0	51.0	0.834	0.90

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STATISTICAL HAZARD ASSESSMENT

To calculate the probability of future volcanic eruptions using a fit distribution as obtained above, we need to consider the residual life distribution of the volcano. This is defined as the probability that an eruption will occur in the next t years, provided no eruptions have occurred in the past x years:

$$S_x(t) = \frac{S(x+t)}{S(x)}.$$

Conversely, the probability that at least one eruption will occur in the coming t years, provided none has occurred in the past x years, is given by $1 - S_x(t)$. This function is directly applicable to the volcano hazard forecast in which x is the number of years since the last eruption, and t is the time in the future for which we want to estimate the eruption likelihood.

In the case of the exponential distribution, $S_{x,exp}$ takes a simple form,

$$S_{x,exp}(t) = \frac{\exp(-\frac{x+t}{\tau})}{\exp(-\frac{x}{\tau})} = \exp\left(-\frac{t}{\tau}\right) = S_{exp}(t),$$

which is independent of the length of the last repose time. For eruption time series that follow a Poisson distribution, the probability of a future eruption therefore does not depend on the time since the last eruption – this is sometimes called the “memoriless property” of the exponential distribution. This is not shared by the other distributions.

The probability for a future eruption is given in Figure 6 and Table 4 for the exponential and log-logistic distributions. While the exponential distribution gives a very high probability of at least one VEI ≥ 2 eruption within the next years (11 % already within one year, and 70 % within 10 years), the log-logistic model provides a much lower prediction (4 % within one year and 29 % in 10 years, respectively). The main reason for this is the length of the last repose period, which strongly influences the predictions of the log-logistic distribution, but remains without effect for the exponential distribution owing to the memoriless property.

Table 4. Prediction of the statistical probability for the occurrence of at least one eruption (VEI ≥ 2) within a given time after 2011

Years after 2011	Probability (exp) [%]	Std. err. (exp) [%]	Probability (log-log) [%]	Std. err. (Log-log) [%]
0	0.00	0.00	0.00	0.00
1	11.31	2.33	3.82	0.75
2	21.34	4.14	7.39	1.43
5	45.13	7.21	16.89	3.08
10	69.89	7.91	29.46	5.89
20	90.93	4.77	46.74	6.59
50	99.75	0.33	71.18	6.89
100	100.00	0.00	85.04	5.36

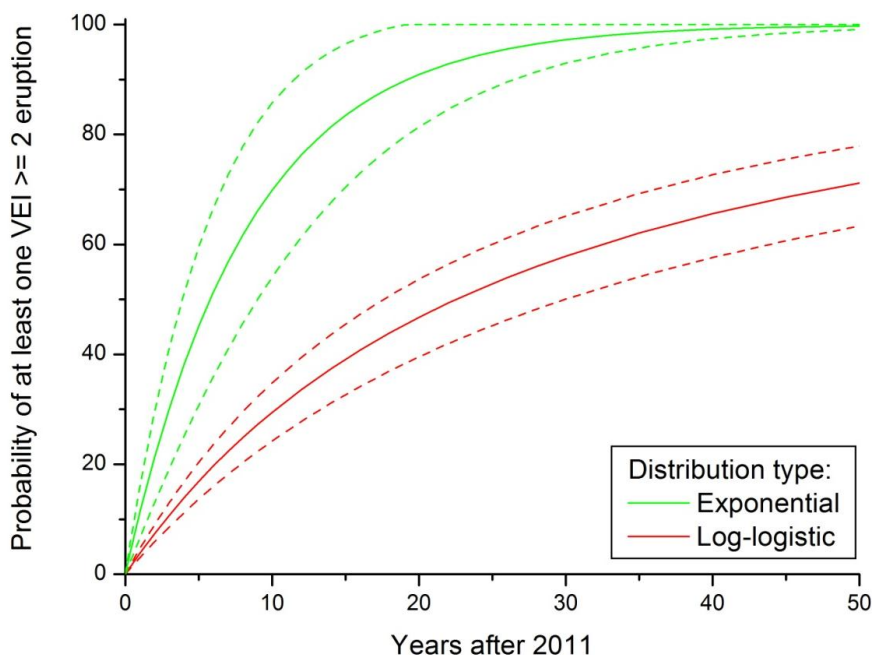


Figure 6. Statistical prediction of eruption hazard, estimated from 2011 on.

Let us consider in more detail the effects of the long last repose time. Taking this last repose period as an anomaly, we can decide to perform the analysis as we would have done in the year 1978, and thereby investigate how likely it was that such a long repose should occur. Changing one repose time within the data set does not significantly influence the fits (Table 3): within one standard deviation, the fit parameters are equal to those from the analysis as performed in 2011. For the memoryless exponential distribution, the difference between the predictions obtained now (Table 4) and those that would have been calculated in the year 1978 (Table 5) are the same within one standard deviation. However, the prediction as it would have been obtained in 1978 counts from 1978 on – i.e., the time axis is shifted with respect to today, and over 95 % probability would have been predicted to witness at least one eruption within 30 years from 1978 – which is, by now. For the log-logistic distribution, in contrast, while again the parameters for an analysis in 1978 are identical to those obtained today within one standard deviation, the length of the last repose time significantly influences the predictions. For the log-logistic distribution, still in 1978 we would have found a very high probability of an eruption within 20 years already (91 %), but since no such eruption has occurred, the predictions calculated today are far decreased.

From this point of view, it appears strange that no eruption has occurred up to today. Yet, this is clearly what has happened. The contradictory predictions found for Puracé certainly present a more striking case than what is normally found in volcano hazard analysis. In most cases, given a good stationarity and independence of the eruption record, the predictions give rise to consistent conclusions well in agreement with the observations; and in fact the problem for Puracé was already noticed at the beginning of the analysis with the anomalous long last repose time. Puracé as a controversial case was chosen to discuss some of the limitations and difficulties of the method, and highlight some of the issues and caveats in the interpretation of the results.

Table 5. Prediction of the statistical probability for the occurrence of at least one eruption (VEI $\geq$ 2) within a given time after 1978, calculated as if the analysis had been performed in 1978

Years after 1978	Probability (exp) [%]	Std. err. (exp) [%]	Probability (log-log) [%]	Std. err. (log-log) [%]
0	0.00	0.00	0	0
1	13.51	2.67	15.11	0.75
2	25.20	4.63	29.32	1.43
5	51.61	7.48	58.27	3.08
10	76.58	7.24	78.84	5.89
20	94.51	3.39	91.18	6.59
50	99.93	0.11	97.64	6.89
100	100.00	0.00	99.17	5.36

SO HOW CAN THIS RESULT BE UNDERSTOOD?

As already mentioned above, it must be admitted that the complications actually come as no surprise: when reconsidering the cumulative number of eruptions over time (Figure 2), clearly the general trend until 1977 was one or two eruptions every decade – even from this simple visual inspection, the long rest after 1977 would appear as unusual. So, is this record somehow “wrong”, and should be reinterpreted, or is it “true”, although statistically not very probable?

Let us first assume that the record is in itself inconsistent from the volcanological point of view. Looking at the eruption record of the last century, eruptions have tended to occur in groups or pairs of several eruptions in consecutive years. If these eruptions had been classed together as one “larger” eruption rather than different events, the record would have looked very different. In fact, we cannot tell without a detailed investigation of the eruption record if the clustering of eruptions in the 20th century presents a real change in volcano behaviour (in the 19th century, no clustering is visible) or just a different interpretation of a constant eruption behaviour, possibly due to different observers, awareness, scientific interest etc. If Puracé is really prone to clustered eruptions, the long last repose time is not so unusual – the repose times between the onsets of the clusters have been of the order of 10-20 years (22 between 1924 and 1946, 10 between 1946 and 1956). In that case, a repose time of 34 years since the last eruption would not be so anomalous. If the clusters recorded from the 20th century were interpreted to represent similar behaviour as the previously observed larger eruptions in the 19th century, we might consider including only large eruptions (here defined as clusters, or VEI = 3) in our record – and possibly also include the single last 1977 eruption, which occurred isolated from all others. In this case, the cumulative number of “large” eruptions (Table 6) shows a rather linear behaviour, with the last repose time hardly standing out from the record (Figure 7). If this were the case, and we were to fit an exponential distribution to this new data set (fitting any distribution more complicated than an exponential to a data set consisting of 7 eruptions can hardly be justified), we would again obtain a good fit (Table 3; in fact, also all the preliminary tests are satisfied). In this case, an analysis in the year 1978 would have predicted a probability of 60 % (with standard error 15 %) of

witnessing at least one eruption within 20 years – here, the large uncertainty makes it reasonably plausible that no such eruption might have occurred yet. Looking from now (as well as from 1978), the probability of future “larger” eruptions is reduced as compared with single $VEI \geq 2$ eruptions, with some 31 % probability of at least one such eruption in the next 10 years (Table 7) – while it is still not well defined what exactly “larger” means for a future eruption.

Although this scenario provides a possible interpretation, it is admittedly speculative and could only be justified if an in-depth investigation of the historical record were undertaken. On the other hand, if the volcano eruption record is really precise as it stands (Table 1), then what do we make of the analysis? We might be inclined to conclude that the volcano behaviour is not stationary enough to warrant the exponential model. In particular, the last long repose time is just slightly out of the 1-standard deviation “limits” around the normal variation in repose times, so maybe the volcano behaviour has just recently changed to a more quiet phase. This hypothesis cannot be confirmed or disproven from a statistical perspective – observatory records, if they existed over a long enough time span, might allow better judgement of this issue. It is true that the exponential distribution does not allow for the volcano to fall into a quiet rest phase (dormancy or extinguishment). The log-logistic distribution can give a better assessment of a volcano that includes both very short and very long repose times due to the interaction of different processes, which is why the predictions differ so much from the exponential model. If the long last repose time is interpreted as a sign of waning activity, then certainly this would lend greater credibility to the log-logistic model than to the exponential, which is inapplicable in such cases.

Table 6. Hypothetical record of “large” eruptions. Included are $VEI \geq 3$ eruptions, clusters of eruptions (eruptions in successive years, dated in the year of the first eruption), and the last eruption in 1977

Start	Stop	VEI	Eruptive characteristics
1847 Oct 27	1852	3 ?	as it was
1869 Oct 4	1869 Nov	3 ?	consider 1869 and 1870 as a single eruption
1885 May 25	?	3	as it was
1924	1927	2++	consider the time between 1924 and 1927 as a single larger eruption (rather than one each year)
1946 Mar	1947	2++	consider 1946 and 1947 as a single larger eruption
1956 (in or before)	?	2++	consider 1956 and 1957 as a single larger eruption
1977 Mar 19	1977 Mar 28 (?)	2	as it was

Given the presumed waning activity of the volcano, one might wonder why the Weibull model does not give a good fit to the data – in theory, the Weibull distribution with shape parameter < 1 should be able to represent waning activity. The reason for this failure is twofold. On the one hand, it is only the last long repose time which makes us speculate that the volcano activity may be diminishing. For fitting the distribution functions, this one repose time cannot influence the fit enough to create a Weibull function with weakening activity (we

have seen above that excluding this eruption from the data set does not significantly change the fit parameters). In our particular example, this gives rise to the observed behaviour of the Weibull fit, which has a shape parameter that cannot – within one standard error – be distinguished from 1, i.e., from an exponential distribution. This is why we decided to abandon the Weibull model for Puracé. If the dwindling activity regime goes on for a longer time, it can be expected that the Weibull model will start to give a better approximation to the data record; until then, we can only rely on the log-logistic distribution for a description of non-Poissonian volcano behaviour.

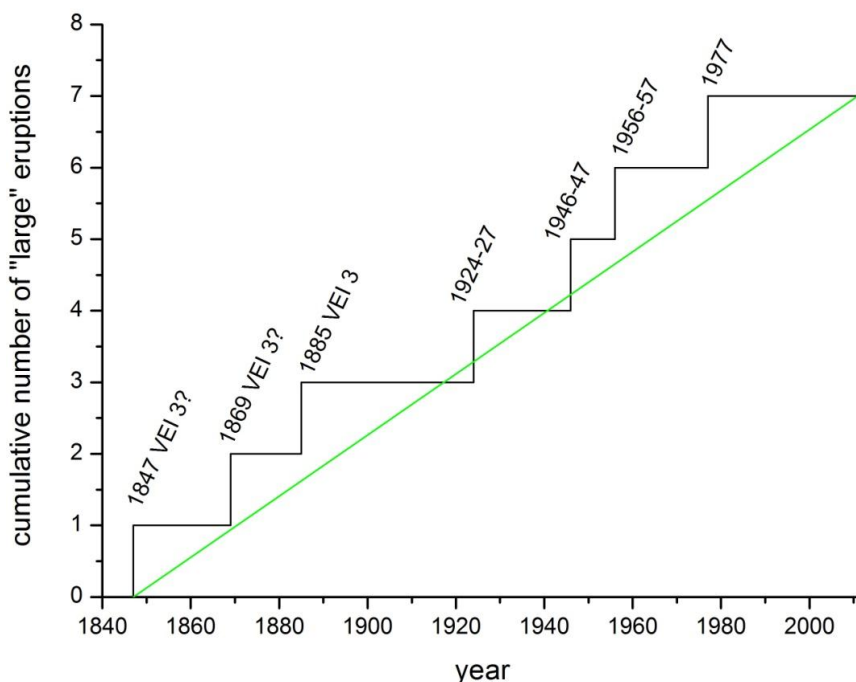


Figure 7. Alternative interpretation of Puracé's eruption record, considering only large ($VEI \geq 3$) eruptions and eruption clusters – cumulative number of eruptions vs. time.

Table 7. Prediction of the statistical probability for the occurrence of at least one "large" eruption ($VEI \geq 3$ or clustered) within a given time after 2011 or 1978 (if the analysis had been performed in 1978)

years after 2011	probability [%]	std. err [%]	years after 1978	probability [%]	std. err. [%]
0	0.00	0.00	0	0.00	0.00
1	3.59	1.43	1	4.49	1.79
2	7.04	2.77	2	8.78	3.42
5	16.69	6.21	5	20.53	7.45
10	30.59	10.34	10	36.85	11.84
20	51.84	14.35	20	60.12	14.96
50	83.89	12.00	50	89.95	9.42
100	97.41	3.87	100	98.99	1.89

Still the log-logistic distribution relies on stationarity of the record. The test for stationarity gave no statistically significant deviations for Puracé, but the last long repose time was slightly anomalous and might, if it continues, indicate a change in volcano behaviour. If fundamentally different levels of activity need to be included in the analysis, one might define the time period until 1977 as one regime of constant activity, and a resting phase since then. The statistical model for the volcano could then be a mixture of exponential distributions with eruption rates corresponding to each regime, and some weighting function for the contribution of each regime in the prediction (Mendoza-Rosas and De la Cruz-Reyna, 2009; Dzierma and Wehrmann, 2010a). More advanced methods might include hidden Markov chains to simulate the switching between regimes (Mulargia and Tinti, 1986; Mulargia et al., 1987; Nathenson et al., 2007; Bebbington, 2007). Both methods are well justified and useful where regimes can be clearly defined either visually (e.g., Nathenson et al., 2007; Mendoza-Rosas and De la Cruz-Reyna, 2008,2009) or mathematically (Mulargia and Tinti, 1986; Mulargia et al., 1987; Bebbington, 2007); for the case of Puracé it appears somewhat arbitrary to distinguish just the last repose time from all others, which is why this approach is not followed here.

As seen from today, there is no necessity of choosing regimes; the eruption record passes the stationarity and independence test and therefore does not preclude the application of the log-logistic or even exponential model. Also, both pass the goodness-of-fit test. So, from the statistical perspective, there is no reason to consider the analysis as incorrect. Given this, we face two questions: Firstly, why has no eruption occurred between 1977 and 2011? And secondly, should we panic at the thought of an immediately impending eruption? The answer to the first question would simply be, statistical analysis can yield improbable results. Seen from 1978, there might have been over 95 % probability of at least one eruption by 2011, but this means that with ca. 5 % probability, no eruption should have occurred. And sometimes the improbable scenario occurs. Consider radioactive decay with a half-life of, say, 7 years. After 30 years, only 5 % of the original radioactive material remains (i.e., every atom out of the starting mass had a 95 % probability of decaying before 30 years were over) – but you will certainly still have the 5 %. You can't say, for each atom, it's just so improbable that it will last 30 years that it should have decayed – no matter how improbable, this is what will happen for 5 % of them... And just like for the exponential function model of the eruption record, these 5 % won't be at a greater risk of decaying than they were 30 years ago. Here is again the memoryless property: after lasting 30 years without decay, each of these nuclides is "as good as new". Again, 95 % of them will decay in the next 30 years, and 5 % of them won't. In the same way, the fact that the improbable situation has arisen that the volcano has not erupted yet does not mean that it must certainly erupt in the near future. The predictions as seen from today have not increased when compared to the hypothetical predictions from 1978. We are still left with a probability of ca. 45 % that at least one ("smaller" VEI = 2) eruption will occur in the next 5 years, if we take the exponential model and assume the volcano is erupting at a constant rate. With the log-logistic model, there is even a chance that the volcano is in a phase of waning activity, which results in decreasing eruption probabilities the further in the past the last eruption occurred. For both models, therefore, the anomalously long last repose time in no way should make us think that an eruption is so long overdue that it must be now imminent; the 1978 analysis is today obsolete and the predictions should be based on today's analysis alone.

For Puracé, this gives us a very wide range of possible eruption probabilities for the near future, between 30 and 70 % within the next 10 years. Somewhat unsatisfactorily, this broad perspective is due to the different predictions from the exponential and log-logistic model. Whilst both models are statistically applicable, we are faced with the decision of which scenario to consider the most plausible.

This is obviously where the statistical method meets its limits – based on mathematics alone, this decision is impossible. Mathematics does allow us to consider the last long repose time as a statistical anomaly, a rare case which does not have any significance at all within a purely exponential behaviour.

Intuitively, I am inclined to disbelieve this and rather think that the eruption record is not exhibiting purely Poissonian behaviour, but instead is better described by a log-logistic distribution with waning activity. If, also plausibly, the clustering of eruptions together with large eruptions were evoked, this would entail in-depth reassessment of the eruption record. In such a situation, it is strongly advisable to substantiate the decision on the statistical conclusions by detailed volcanological data that provide insight into the physical processes occurring in the volcanic system.

CONCLUSION

The historical eruption record of Puracé Volcano, while relatively well documented over almost two centuries, presents a recent anomaly, with no eruption since 1977. This long last repose time, being the only such deviation from a near-constant eruption rate, does not influence the statistical properties of the record strongly enough to preclude a statistical hazard analysis.

Stationarity and independence tests give good results in spite of the long last repose period, and the fit functions to the survival function pass the goodness-of-fit tests. However, the different response of the exponential and log-logistic model to the last repose time results in very different predictions for future eruption probabilities.

If the correctness of the record is taken for granted, the most plausible explanation seems to be that the log-logistic, rather than exponential, model gives a good description of Puracé's recent behaviour. In this case, we would face a ca. 17 % probability of an eruption in the next 5 years, rising to ca. 30 % within 10 years. However, a change in the record from the occurrence of several larger (VEI = 3) eruptions in the 19th century to clustered eruptions in the 20th century may be indicative of a change in how eruptions were classified in the record and warrants further research into the completeness and accuracy of the record, which may also shed more light on the interpretation of the latest period of repose and influence the outcome statistical hazard analysis.

This analysis is a contribution to hazard assessment on a longer time scale than can be achieved by monitoring efforts, which it is meant to complement. While not providing warning of immediate impending volcanic hazards like monitoring can, this long-term hazard assessment may serve for land planning purposes and risk awareness, and can be performed even without monitoring data. It is thus useful to estimate the likely hazard from a volcano on the basis of its historical behaviour even before the possible reactivation of the system.

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Chapter 7

**RESOLUTION OF DIFFERENT-SCALE DEFORMATION
DATA MEASURED AT VOLCANOES AND ITS
IMPORTANCE FOR ERUPTION MONITORING:
VOLCÁN DE COLIMA, MÉXICO**

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ABSTRACT

The measurements of deformation at volcanoes may be carried out by ground-based techniques (measurements of tilts and distances) and by remote sensing technique (e.g., interferometric synthetic aperture radar InSAR). This review of the different-scale tilt changes and InSAR measurements carried out at Volcán de Colima, México during the 2004-2011 eruptive activity shows the ability of the deformation measurements with different level of spatial and temporal resolution to give different type of information about the volcanic process.

The minute-scale tilt changes were able to describe the individual explosions, the daily-averaged tilt changes indicated the eruptive episodes, and monthly averaged tilt changes showed the general tendency in deformation process but indicating also the eruptive episodes which were associated with inflationary-deflationary cycles. InSAR measurements, in their turn, allowed to recognize the general tendency in deformation process at volcano (continuous subsidence) but not indicated any influence of two extrusions that occurred during the period of observations. It is shown that a complex study of the deformation process at volcano including different-resolution techniques would be informative during volcano eruption monitoring.

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1. INTRODUCTION

Study of deformation of volcanic edifice and its surrounding at different stages of volcanic activity is an important tool for volcano activity monitoring (Dzurisin, 2007). The measurements of deformation may be carried out by ground-based techniques (measurements of tilts and distances) and by remote sensing technique (e.g., interferometric synthetic aperture radar InSAR). Each of these techniques has their advantages and specific features. Our review presents the comparative analysis of the resolution of the InSAR survey and different-scale tilt changes for monitoring the surface manifestations of eruption process. These both techniques for surface deformation studies are widely used in volcanic investigations. InSAR technique allows detecting deformation extending over large areas without requirement for field instrumentation. It was used to study a rifting episode in Afar, with a 65 km (more than 1 km³) long intrusion (Wright et al., 2006; Grandin et al., 2009), as well as large accumulations of magma at depth (Froger et al., 2007; Ruch et al., 2008).

The InSAR survey, covering about 900 remote volcanoes of the central Andes, revealed broad roughly axi-symmetric deformation at some volcanic centers: two stratovolcanoes were inflating (Uturuncu, Bolivia, and Hualca Hualca, Peru) and a caldera Cerro Blanco in northwest Argentina was subsiding (Pritchard and Simons, 2004).

Main interest of InSAR data is in their spatial coverage however it is compensated by a rather poor temporal resolution with a revisit time around one month. A large spatial coverage potentially enables to detect deep sources of deformation. A tiltmeter is an instrument designed to measure very small inclination of the ground. Long-term tiltmeter records at Kilauea Volcano, Hawaii show that basaltic magma moving beneath the summit caldera of the volcano forces the ground to move up and down, sometimes by as much as 2 m. A continuously (November 1982 to July 1986) record of tilt from the Uwekahuna tiltmeter located on rim of Kilauea's caldera shows many peaks indicating an increase and decrease in slope angle which indicates the summit area was inflating and deflating with basaltic magma. The duration of these peaks may be of about one month (Johnson, 1992).

The continuous (beginning from 1977) tilt monitoring at basaltic Mt. Etna showed that the tilt changes have accompanied the main eruptive phases of the volcano including the 1991-1993 opening of eruptive fissure and the 1989 propagating 7-km-long fracture (Bonaccorso et al., 2004). Very small tilt changes (order of 0.1-1.5 microrad) were detected during the explosive episodes in 1995-2000 and the 1998-2000 lava fountains (Bonaccorso, 2006). Rather low sample rate (48 counts / day) did not allow recognizing the details of tilt changes during the explosions indicating the general reaction of instruments only.

Studying daily averaged tilt changes, Chadwick et al. (1988) showed that six lava extrusions at Mount St. Helens, occurring during May 1981 to August 1982, were preceded by inflationary process. The instruments installed within 50 m of the dome (250-300 m from the vent) generally recorded a total of 2000-4000 μ rad of tilting. Outward tilting began several weeks before each extrusion, accelerating sharply for several days and then changed direction to inward tilting before eruptive activity. Four-day-long inflationary tilt changes were observed before lava extrusion at Pinatubo volcano on 7 June 1991 (Ewert et al. 1996). Bonaccorso et al. (2008) observed that the inflationary tilt change during the first days of the February-March 2007 Stromboli eruption was in good correspondence with the lava effusion rate. Tilt change monitoring during the 1996-1997 eruptions at andesitic Soufriere Hills

volcano, Montserrat revealed 6-14 hrs inflation cycles supposedly associated with the magma pressurization at shallow depths (Voight et al., 1998).

The cited publications described every time the application of one only method or one only technology of the signal processing. Our article demonstrates the ability of deformation information of different scale resolution to indicate various aspects of the same volcanic eruption.

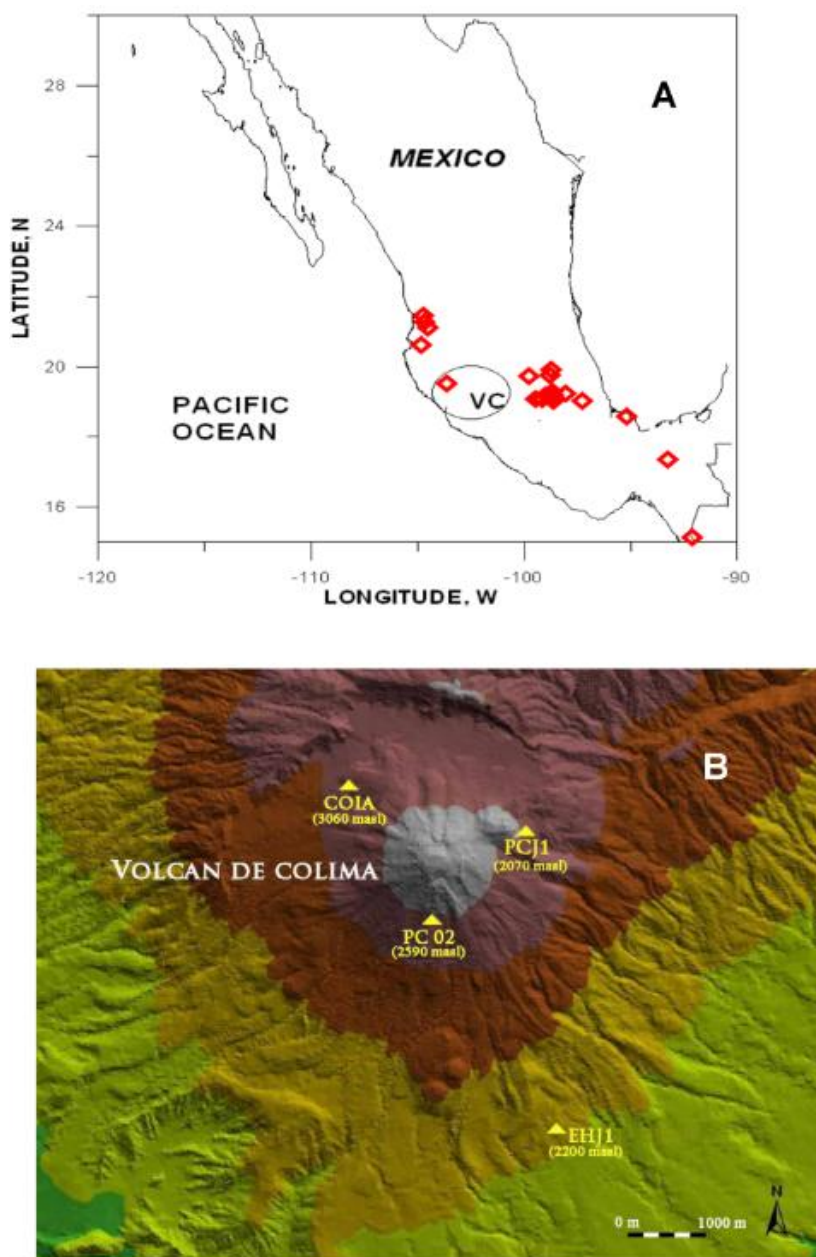


Figure 1. Position of Volcán de Colima within Mexican Volcanic belt (A) and the tilt network around Volcán de Colima (B). Diamonds in A show the active volcanoes; circle shows the position of Volcán de Colima (VC). Triangles in B show the tilt stations.

Volcán de Colima, the most active Mexican volcano, whose recent unrest began from 1997, was chosen as an example for this comparative analysis.

Volcán de Colima is an andesitic stratovolcano reaching 3860 m above sea level (a.s.l.). It belongs to the Trans-Mexican volcanic belt (Figure 1). The recent 1997-2012 unrest of the volcano was characterized by four extrusion episodes with lava dome building in the crater in 1998-1999, 2001-2003, 2004 and 2007-2012 (Zobin et al., 2002; 2011; Ramírez-Ruiz et al., 2002). In this article, we analyze the resolution of different techniques that were applied for study of deformation of volcanic edifice associated with the 2003-2011 period in eruptive activity at Volcán de Colima.

During this period of activity, two effusive eruptions of May 2001-February 2003 and September-October 2004 produced $6.3 \times 10^6 \text{ m}^3$ and $7.7 \times 10^6 \text{ m}^3$ of lava, respectively, with the building of the $2 \times 10^6 \text{ m}^3$ lava domes in the summit crater and generating lava flows reaching 1.4-2.4 km of length (Zobin et al., 2008).

The lava domes were destroyed by the sequences of Vulcanian explosions during July-August 2003 and March-September 2005. The recent lava dome began to grow in February 2007 and filled the summit crater in June 2011 reaching volume of $2 \times 10^6 \text{ m}^3$. Characteristic manifestations of eruption activity, such as Vulcanian explosions and lava dome, are shown in photos of Figure 2.

2. INSAR CHARACTERISTICS OF THE DISPLACEMENT FIELD FOR THE REGION OF VOLCAN DE COLIMA

Pinel et al. (2011) used the time series of SAR data acquired from December 2002 to August 2006 (Figure 3A). This period included the February 2002-February 2003 and the September-October 2004 lava extrusions. Two descending and one ascending tracks were used. On the descending tracks the satellite looked toward the west; on the ascending track, satellite looked toward the east. The minimum time delay between successive data was 35 days. Specific acquisition conditions characterizing andesitic stratovolcanoes, mainly vegetated areas with large elevation ranges, induced low signal coherence as well as strong tropospheric artifacts that resulted in small signal-to-noise ratio.

Figure 3B shows the residual mean velocity in line of sight (LOS) measured for the summit area of Volcán de Colima by Pinel et al. (2011) along the tracks shown in Figure 3A. Pinel et al. (2011) concluded that for the summit area of Colima Volcano a displacement away from the satellite was clearly observed on the three tracks.

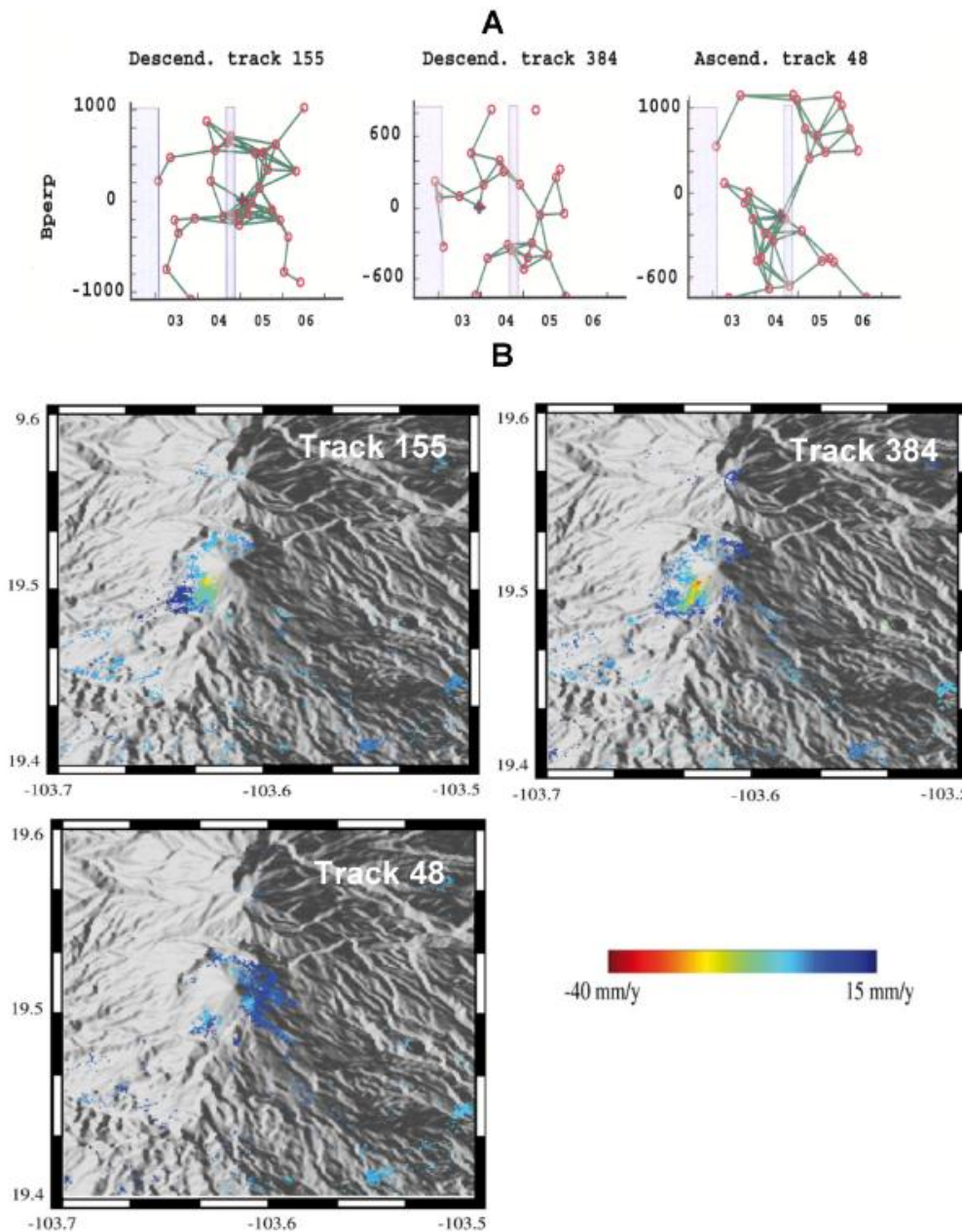
This displacement was considered as corresponding to a subsidence of up to 30 mm/year. Analysis of LOS in of the volcano showed that the subsidence was centered on the summit of Volcán de Colima and decreased away from the summit (from 0.45 to 2 km) corresponding to a tilt around 6–9 μrad . No correlation was obtained between the presence of a dome of magma within the summit crater and the observed subsidence for instance. The maximum subsidence was located by Pinel et al. (2011) on the lava flow emplaced in 1998 on the southwest flank of the volcano summit. It may be caused by significantly greater output of magma during the 1998-1999 extrusion ($39 \times 10^6 \text{ m}^3$) comparing with the magma output during the effusive eruptions of 2001-2003 ($6.3 \times 10^6 \text{ m}^3$) and 2004 ($7.7 \times 10^6 \text{ m}^3$) (Zobin et al.,

2008). A summit deflation was measured from 1 to 3 km around the summit of Colima Volcano. No inflation before the September 2004 lava extrusion was recognized.



Photos were taken by Carlos Navarro and Mauricio Bretón González.

Figure 2. Examples of the 2004-2012 explosive (A, Vulcanian explosion of 8 December 2004) and effusive (B, Lava dome in the crater of volcano, 21 April 2009) activity at Volcán de Colima.



Modified from Pinel et al. (2011).

Figure 3. Perpendicular (Bperp) versus temporal baseline plot of the ENVISAT data for three tracks 155, 384, and 48 (A) and the residual mean velocity in LOS in mm/year for the summit area of Volcán de Colima obtained for these three tracks (B). There were used 30 images and 98 interferograms for descending track 155, 26 images and 54 interferograms for descending track 384, and 30 images and 89 interferograms for ascending track 48. In A, circles represent the images used in the PS processing, and lines represent the interferograms formed in the small baseline approach. Shaded grey areas correspond to periods of effusive activity at Volcán de Colima. In B, positive values are for displacements directed towards the satellite. VdeC is Volcán de Colima. The model used considers both a uniform circular load of $1.3 \text{ Mm}^3/\text{y}$ with a 0.5 km radius and a Mogi deflating source at 2.5 km depth with $\Delta V=0.23 \times 10^6 \text{ m}^3/\text{y}$.

No deflating signal was associated with the emplacement of the 2004 effusive products: the dome as well as lava flows. Two 2004 lava flows (the lengths and widths of 2400 m and 300 m respectively) emplaced on the N and WNW slopes of Volcán de Colima were not imaged by the selected pixels presumably due to decorrelation induced by the flow emplacement. A $2 \times 10^6 \text{ m}^3$ 2004 lava dome was emplaced in the summit crater. Once again, as noted Pinel et al. (2011), the displacement induced by the dome load was too small to account for the observed subsidence, which is not restricted to periods where a dome was present within the crater. Pinel et al. (2011) did not observe any variation in the subsidence rate over the studied time period despite of the 2004 lava extrusion. Modeling of a small spherical expansion Mogi source (Mogi, 1958) embedded in a homogeneous half-space is widely used to simulate ground deformation produced by local perturbation like volcanic magma chamber. The best Mogi model for Volcán de Colima was calculated by Pinel et al. (2011) using the three tracks shown in Figure 3A, considering displacement recorded at less than 5 km from the summit of Colima Volcano and masking the area corresponding to the 1998 lava flow. The best solution was obtained considering a deflating source located at 2.5 km depth beneath the crater with a volume decrease of $0.23 \times 10^6 \text{ m}^3/\text{y}$.

3. TILT CHANGES AT VOLCAN DE COLIMA DURING THE 2004-2010 PERIOD

The near-field ground-based deformation measurements at Volcán de Colima were represented by tilt observations. We show the comparative resolution of the elements of eruption activity using monthly averaged, daily averaged and minute-scaled tilt changes.

3.1. System of Tilt Monitoring

The tilt observations at Volcán de Colima began in April 1999 when the first station COIA was installed at a distance of 1.6 km (elevation 3060 m a.s.l.) from the central crater (Figure 1B). Additional three stations were installed in October 2002 (EHJ1; 4 km from the crater, elevation 2200 m a.s.l) and in May and June 2006 (PC02; 2.4 km from the crater, elevation 2600 m a.s.l., and PCJ1; 1.5 km from the crater, elevation 3200 m a.s.l.). The stations were equipped with the Applied Geomechanics 701-2 biaxial tiltmeters deployed in 1-m-deep boreholes at monitoring sites protected against the influence of climatic factors (temperature, humidity and corrosion factors). These resistive bubble-type tiltmeters have a resolution of 0.1 μrad , a rate of 30 samples/hour, and included a temperature sensor sealed with the sensor body.

The radial components of both tiltmeters were oriented towards the crater; so positive signals indicated inflation whereas negative signals indicated deflation. The sampled data of the registered components (tangential and radial) as well as the temperature recorded at the site were transmitted to the monitoring center in Colima using a system of data logger and Freewave type transmitter and were corrected for the temperature and humidity factors. Temperature compensation of Applied Geomechanics tiltmeter readings was performed using equation

$$\theta = SV - KZx(T - T_{cal}) \quad (1)$$

where θ = true angular position (tilt);

T = the temperature at which your measurement was made;

T_{cal} = the calibration temperature reported in the tiltmeter user's manual;

S = scale factor at temperature T ;

V = the measured output voltage at temperature T ;

KZx = calibration constant.

The tilt stations are very sensitive to external influences; the abrupt drift of tilt components may be related to intense rainfall or local earthquakes. During the period of observations, from 2004 to 2010, no significant ($M \geq 5$) local earthquakes were observed. No significant effects of the rainfalls were observed. All significant variations in tilt change during the period under study can be attributed to the deformation of the volcanic edifice.

3.2. Minute-Scaled Tilt Changes: Recognition of Single Volcanic Explosions

The recent monitoring of Volcán de Colima showed that the tiltmeter with the sample rate of 30 counts / hour may record rather large (order of up to 100-150 μrad) tilt changes with few-minute wavelengths associated with the Vulcanian explosions.

These few-minute impulses appear on the tilt records of near stations when the explosion is large enough to generate an explosive shock corresponding to the instrument sensibility. Zobin et al. (2007) had examined the tilt records coincided with the appearance of 120 explosions that were observed during the March-September 2005 explosive sequence. The impulses of tilt were recognized for 46 events only. The remaining 74 explosions had no response on the tiltmeter records or were recorded with missing of the initial part of signal.

Analysis of the energetic level of these 120 explosions (energy of these explosions was estimated from broadband seismic records in Zobin et al., 2006) showed that the large-amplitude tilt changes were recorded only for 17% of the explosions with energy of 10^7 to 10^8 J but for 51% of the explosions with energy of 10^9 to 10^{10} J and for 87% of the explosions with energy of 10^{11} to 10^{13} J.

The energy threshold of 10^9 to 10^{10} J is supposed to be limited by the instrument sensibility. The recording of impulses produced by the explosions with energy lower than 10^9 to 10^{10} J may be conditioned by their closer position to the instrument site. Figure 4 shows the typical tilt changes recorded at the station COIA (radial and tangential components; distance of 1.6 km from the crater) and associated with Vulcanian explosion of 4 January 2005. The explosion was identified with simultaneous video image (Figure 4A).

The tilt changes associated with Vulcanian explosions formed the bi-modal or one-modal impulses of duration of 3 to 6 min. The radial component of impulses had positive polarity (away from the crater) for all events; the tangential components had both positive and negative polarities.

The difference in polarities in the tangential components and their large amplitudes indicated that the position of the explosion sources was not permanent and did not coincided with the formal direction from the instrument site to the crater.

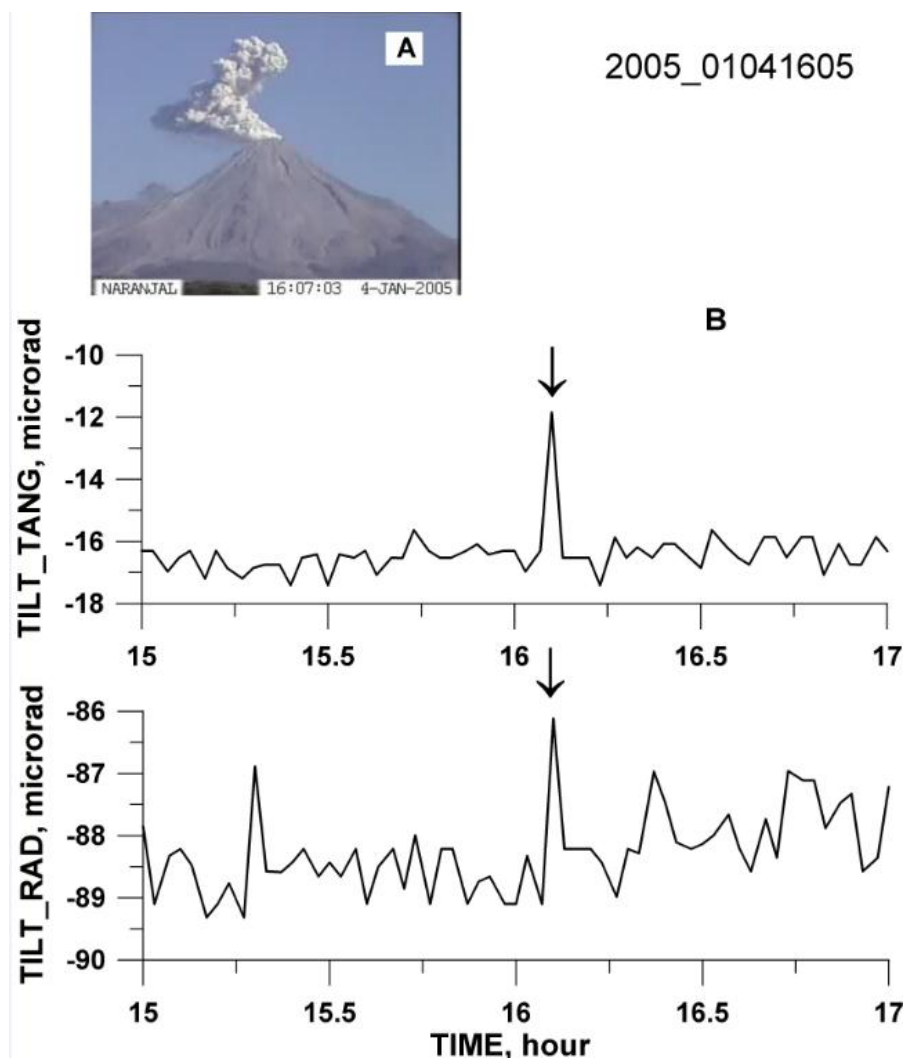


Figure 4. (A) Vulcanian explosion at Volcán de Colima recorded on January 4, 2005 at 16:05 GMT (code 2005_01041605) by a video station NAR, situated at a distance of 15 km from the crater, and (B) the two-component two-hour tilt records at station COIA situated at a distance of 1.6 km from the crater. There are seen the clear impulses on the both components of tilt record (marked by arrows) coincident with the moment of the explosion.

The relatively low sampling rate of the tilt records (1 sample/120 s) does not allow direct comparing with the simultaneous sequence of video images of the explosions. Therefore, it was impossible to say if the tilt changes began before the surface explosion manifestation or during this process.

3.3. Daily Averaged Tilt Changes: Recognition of the Eruption Episodes

Continuous tilt changes during the 2004-2005 effusive-explosive episodes at Volcán de Colima (México) were recorded simultaneously by two tiltmeters COIA and EHJ1 that were

installed on opposite sides of the volcano at elevations of 3060 m and 2200 m above sea level (a.s.l.). This stage of the 1997-2012 unrest at Volcán de Colima began on 30 September 2004. The extrusion of andesitic lava during September – October 2004 formed a summit lava dome. A series of large Vulcanian explosions (with energy $\geq 10^{11}$ J) began on 10 March and continued up to 27 September 2005. The March-September 2005 explosive sequence removed the 2004 lava dome and left a crater 260 m wide and 30 m deep. The ejected material had a volume of about $2 \times 10^6 \text{ m}^3$ (Zobin et al., 2011).

Analysis of daily averaged tilt changes allowed to discriminate the anomalies associated with different stages of eruption activity (Zobin et al., 2011). Figure 5B shows a symmetrical inflationary-deflationary impulse, preceding the 30 September lava extrusion, that was recorded at the nearest COIA station. The impulse began to grow on 6 September, reached its maximum (6 μrad) on 10 September and was terminated 5 days before the lava began to flow from the crater. Station EHJ1 did not record any anomaly.

During the lava extrusion, the tilt record at station COIA (Figure 5B) showed a well-expressed negative impulse, which began on 5 October, 6 days after the beginning of the extrusion, indicating deflation of the upper part of volcanic edifice at an elevation of about 3000 m. This symmetrical deflationary-inflationary impulse had duration of about 170 hours and reached an amplitude of -8 μrad on 9 October. The peak in the tilt deflationary impulse occurred just after the lava effusion rate reached its maximum (Zobin et al., 2011). It occurred against a background of a flat record that began simultaneously with lava extrusion from the crater on 30 September. The flat record re-appeared after the deflationary impulse and then continued until 23 October, marking the end of lava effusion. No significant tilt change at station EHJ1, situated at an elevation of 2200 m on the southern side of the volcano, was recorded before or during the lava extrusion (Figure 5C).

Immediately after termination of lava discharge, the tilt changes recorded at upper station COIA began to show an intense inflation, which continued for three months (Figure 5B; stage II). Tilt amplitude reached about 50 μrad . The record of inflation at lower station EHJ1 began about 50 days later and was characterized by amplitude of about 20 μrad only.

The initial stage of inflation was accompanied by numerous small degassing events (Figure 4a). Their number increased simultaneously with the amplitude of the inflation anomaly at station COIA (elevation 3060 m) but then gradually decreased when the inflation at station EHJ1 (elevation 2200 m) began.

The stage, during which there were large explosions with energy $\geq 10^{12}$ J, was divided into two substages: the large explosions that occurred in March, 2005 and during May-June, 2005. The energy of explosions was estimated from the broadband seismic records in (Zobin et al. 2006; 2007). During the March 2005 explosion sequence, upper station COIA (Figure 5B) recorded a deflationary-inflationary anomaly which began on 9 February and continued up to 7 April with the maximum apparent deflation amplitude of about -25 μrad recorded on 18 March. Station EHJ1 recorded a small-deflationary tendency reaching on 18 March about -4 μrad only. The large explosions occurred on 10 and 13 March during the maximum development of a deflationary anomaly.

During the May-June 2005 explosion sequence, upper station COIA recorded rather low-amplitude variations in tilt while lower station EHJ1 recorded a well-expressed deflationary anomaly from 15 April to 3 July with the apparent deflation amplitude of about -76 μrad on 5 June.

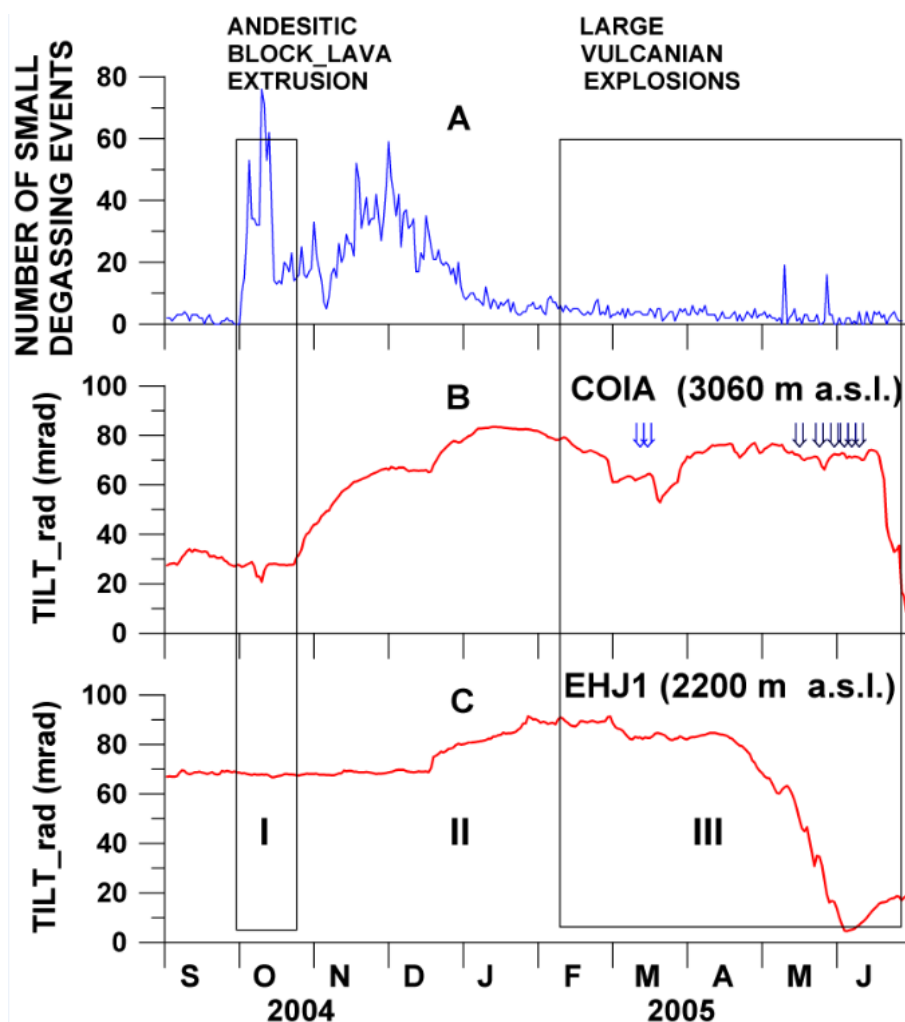


Figure 5. (A) Daily number of degassing events. (B and C) The daily averages of tilt data (radial components) recorded by two tilt stations COIA and EHJ1 situated at different elevations on the volcanic edifice during September 2004 – June 2005 and corrected for temperature effects. Stage I locates the tilt changes during the andesitic blocky-lava extrusion; stage II shows the period of transition from effusive to explosive stages, and stage III shows the period of large Vulcanian explosions. The number of degassing events was calculated from the seismic records at short-period station EZV4 situated at 1.6 km from the crater. Arrows show the moments of large explosions observed in March 2005 and during May-July 2005.

The anomaly began about 30 days before the first large explosion of this sequence. All May-June large explosions occurred during the development of deflationary anomalies. The maximum deflation was correlated with the largest explosion of this sequence which occurred on 5 June, with energy of about 1.5×10^{13} J. Two deformation sources were located by Zobin et al. (2011) based on the tilt amplitude ratios.

The first was situated at a depth between 300 m and 1800 m beneath the crater at the northern flank of the volcano and was responsible for volcano deformation during the preliminary September 2004 stage, the October 2004 extrusion, the initial stage of the transition period, and the March 2005 explosion sequence. The second source was located at a

depth between 1800 m and 2800 m beneath the crater at the southern flank of the volcano and was responsible for volcano deformation.

3.4. Monthly Averaged Tilt Changes: Recognition of the Inflationary-Deflationary Cycles

Figure 6 shows monthly averaged tilt changes (radial component) observed at four tilt stations of the network during the 2004-2011 period. This period includes the September-October 2004 extrusion with a building of the lava dome, the March-September 2005 sequence of explosions destructing this lava dome, and the recent lava dome building that began in February 2007 and continued till 2012. The curves of cumulative volume of lava, erupted during 2004-2011, are shown in Figure 6A.

Monthly averaged tilt changes at four stations are characterized by different behaviour (Figure 6B). Two stations, COIA and EHJ1 that were in operation during all this period showed rather similar tendencies.

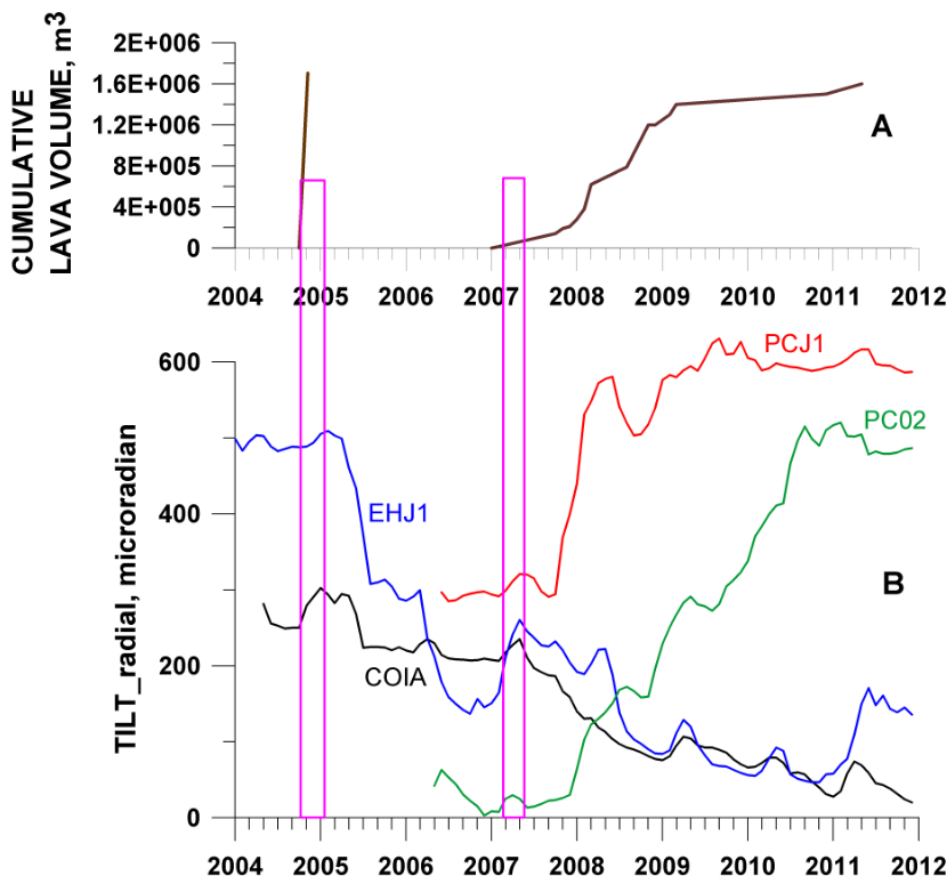


Figure 6. Cumulative lava volume erupted during two 2004-2011 effusive episodes at Volcán de Colima (A) and monthly averaged tilt changes recorded at four tilt stations (B). The periods of inflation stages associated with the 2004 and 2007-2011 effusive eruptions are shown in rectangles.

They demonstrate the permanent subsidence of the region during 2004-2011 that coincides with the InSAR observations carried out by Pinel et al. (2011). At the same time, monthly averaged tilt changes allow to resolve two similar inflationary-deflationary stages associated with lava extrusions. Their inflationary stages are shown in rectangles in Figure 6.

The 2004 lava extrusion was associated with the inflationary-deflationary impulses that were observed at both stations after the cease of lava effusion.

The maximum tilt change of the inflationary stage, which continued for about 4 months, reached 50 μrad at station COIA and 20 μrad at station EHJ1. Beginning from February 2005, the continuous deflation was observed at both stations.

The 2007-2011 lava dome building was associated with the inflationary-deflationary impulses that were observed at both stations during the initial stage of lava effusion. The maximum tilt change of the inflationary stage, which continued for about 4 months, reached 30 μrad at station COIA and 50 μrad at station EHJ1. Beginning from May 2007, the continuous deflation was observed at both stations.

Stations PCJ1 and PC02 that were installed later, in May-June 2006, demonstrated also the inflationary-deflationary impulses that were observed during the initial stage of lava effusion in February-May 2007. The maximum tilt change of the inflationary stage, which continued for about 4 months, reached 30 μrad at station PCJ1 and 15 μrad at station PC01. At the same time, these stations, in contrast with stations COIA and EHJ1, recorded later on continuous inflation. These stations are situated near one another forming a group (See Figure 1B). It is possible to suggest the local effect of sites or local magma movement just beneath the station zone. Therefore, monthly averaged tilt changes may be sensitive to the eruptive episodes but not to the details of the eruption. The inflationary-deflationary impulses appeared just after the cease of the short-term 2004 effusive activity but during the initial stage of the long-term 2007-2011 effusive activity. General decrease of tilt changes with a distance from the crater during the inflationary stages indicates a position of the source of deformation beneath the crater of the volcano.

4. DISCUSSION AND RESULTS

Recent review of deformation process measurements at Volcán de Colima, carried out by different methods and with different level of data smoothing, illustrated the ability of each of them to demonstrate different aspects of the eruptive process. This review shows that a complex study of the deformation process at volcano including different-resolution techniques would be informative during volcano eruption monitoring.

The InSAR large-scale observations did not allowed to see the influence of recent lava extrusions on the total field of deformation but allowed to locate a long-lived source of deformations beneath the volcano edifice of Volcán de Colima and to recognize the general tendency in deformation process.

The minute-scale tilt observations allowed to recognize the tilt changes associated with an explosion at volcano. The daily averaged tilt changes are good associated with the surface manifestations of volcanic activity. The monthly averaged tilt changes are supposed to be more dependent on the site of installation. This scale tilt changes allow to see any temporal

associations with eruption episodes but they are not good correlated with the surface manifestations of the eruption.

The deformation sources calculated by Pinel et al. (2011) and Zobin et al. (2011) from the InSAR observations and the daily averaged tilt changes, respectively, are shown in Figure 7. They are in a good accordance. Daily averaged short-scale tilt changes showed the presence of at least two local sources of deformation that appeared during magma migration associated with effusive eruption. One of these local sources (deep source) was proposed at the same depth as the long-lived source of deformation while the surface source of deformation, obtained from daily averaged tilt changes, may be associated directly with the lava extrusion.

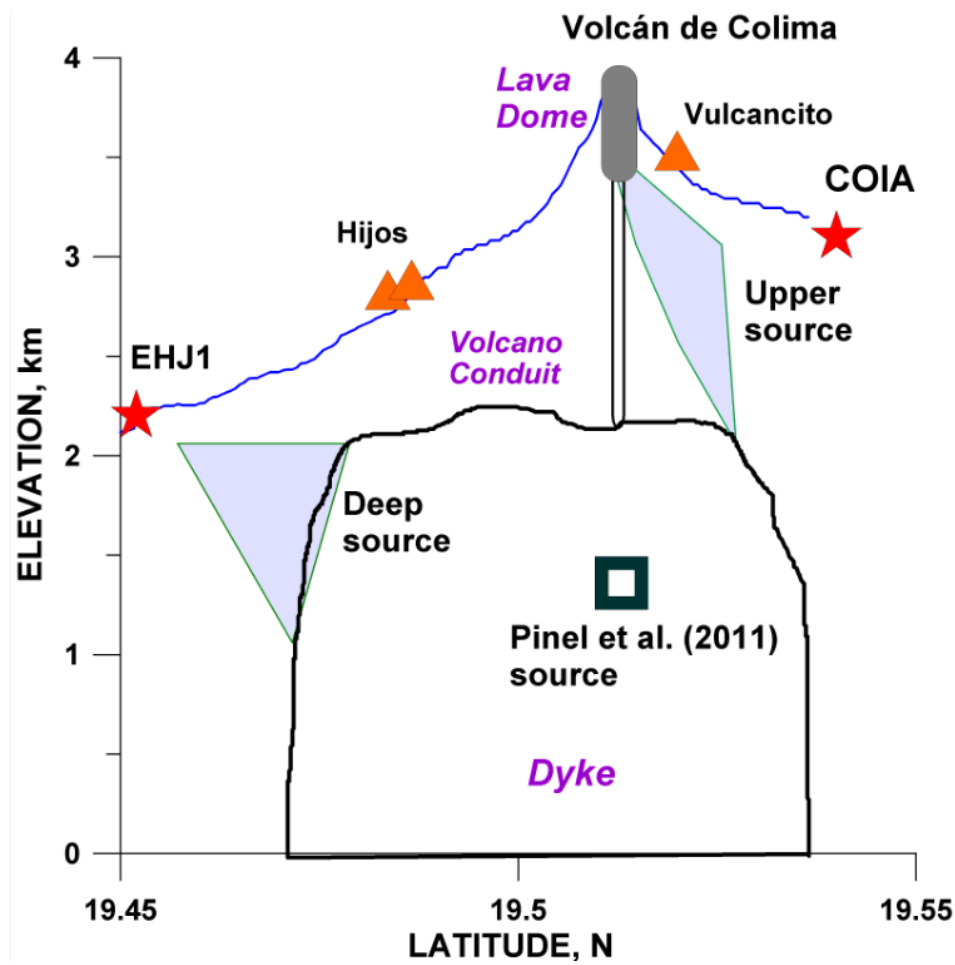


Figure 7. Approximate positions of the deformation sources during the 2004-2005 eruption at Volcán de Colima shown on a north-south profile of Volcán de Colima and the model of volcanic system. The schematic view of the volcanic system consisting of the dyke conduit and lava dome is shown. The source areas correspond to the position of the sources that were constrained by the minimum of the misfit function of the tilt amplitude ratios in Zobin et al. (2011). The position of the deformation source calculated by Pinel et al. (2011) from InSAR observations is shown as a square.

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Chapter 8

THE 2007-2012 LAVA DOME GROWTH IN THE CRATER OF VOLCÁN DE COLIMA, MÉXICO, DERIVED FROM VIDEO MONITORING SYSTEM

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ABSTRACT

Monitoring lava dome growth is difficult because of dangerous access and repeated destruction during explosive episodes. The Video Monitoring System (VMS) installed at Volcán de Colima, México, allowed us continuous monitoring of the lava dome growth in the crater of the volcano during the period of 2007-2012. The lava dome growth developed with the mean rate of 0.26 m/month until it reached its maximum height of about 84 m in its central zone in January 2010 and filled the crater. Beginning from February 2010, the partial collapses began in the western zone of the dome. The construction of the 3-D model of the dome gave us a possibility to calculate the volume ($2.69 \times 10^6 \text{ m}^3$) and the area ($1.19 \times 10^6 \text{ m}^2$) of the 2010 lava dome. The methodology of video monitoring of a volcano developed in this Chapter may be recommended for application at other volcanoes.

Keywords: Terrestrial photogrammetry, Georeferencing, Volcán de Colima

1. INTRODUCTION

Lava dome growth at different volcanoes of the world have been the subject of many investigations beginning from the famous “Mimatsu-diagrams”, illustrating the growth of the Showa-Shinzan lava dome of Usu, Hokkaido, volcano during 1943-1945 (Minakami, 1951). This study was conducted by Masao Mimatsu by drawing the outline of the dome on a

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transparency with a theodolite installed at a fixed place near the growing dome. Recent studies of the lava dome growth at Mount St. Helens (Major, 2009; Schilling et al., 2008), Soufrière Hills volcano, Montserrat (Ryan et al. 2010; Wadge et al. 2010; Hale, 2009), Shiveluch volcano, Kamchatka (Ramsey et al., 2012), and Redoubt volcano, Alaska (Diefenbach, 2012), among others, were carried out using modern video, time lapse imaging aerophotogrammetry, and thermal infrared remote sensing techniques. These studies allowed estimates of the rates of lava dome growth and the volumes of growing domes to be made, and allowed the modeling of volcanic edifices through time.

Our Chapter discusses the methodology and results of continuous video observations of the 2007-2012 lava dome growth in the crater of Volcán de Colima.

Volcán de Colima, the most active volcano in México, is an andesitic stratovolcano with a maximum height of approximately 3860 m above sea level (a.s.l.). Together with the dormant volcano Nevado de Colima (4,330 m a.s.l.) they represent Colima Volcanic complex within the Transmexican Volcanic Belt. Volcán de Colima is located 32 km north of the city of Colima (Figure 1).

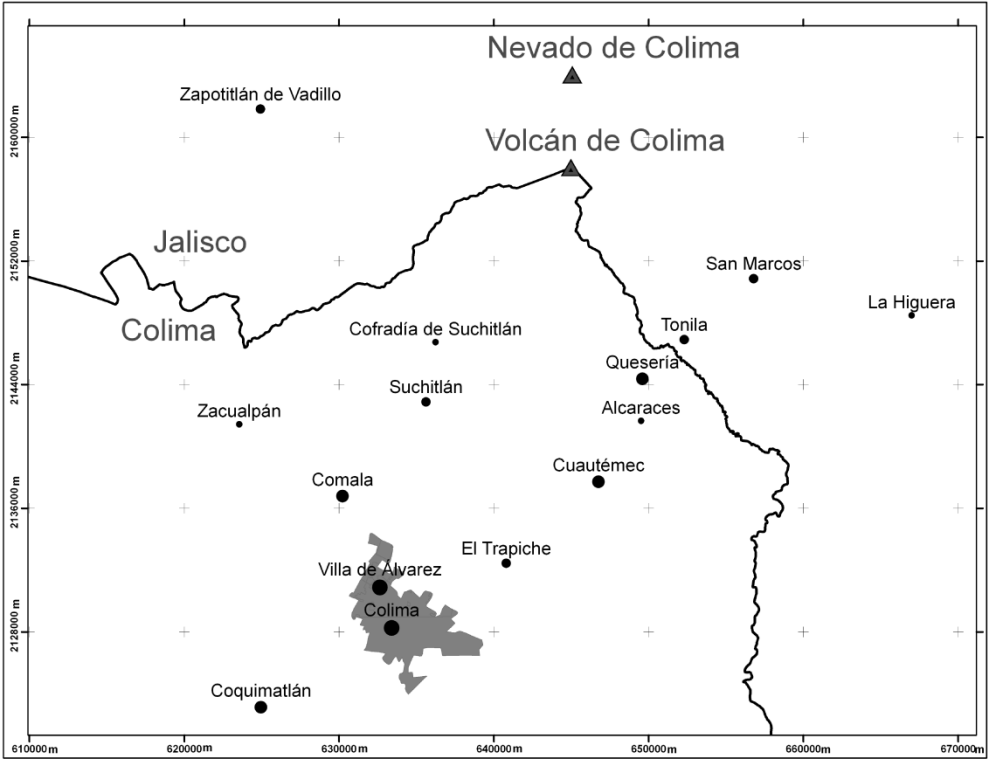


Figure 1. Position of Volcán de Colima and the nearest populations shown as filled circles. The volcanoes are shown with triangles.

Over the last 500 years the Volcán de Colima has had more than 40 significant eruptive events (Bretón et al., 2002). The process of lava dome building and subsequent lava dome destruction in the crater of the volcano are reoccurring events. This process was firstly documented in a publication by Bárcena (1887), in which he wrote about the lava dome building accompanied by the lava flow along the SW slope of the volcano during January

1886. Breton (2011) wrote about the lava domes which filled the crater of Volcán de Colima for the years 1901, 1905, between 1930 and 1931, 1937, 1941, 1943, 1951, 1957; and during the 1958-61, 1966-1967, 1975-1976, and in February 1981. The lava dome that was built between February and March of 1991 was destroyed on 21 July 1994 by an explosion which left a crater with a diameter of 140 m and a depth of 50 m (Bretón, 2011).

The recent 1997-2012 unrest of the volcano was characterized by four episodes of extrusion with lava dome growth in the crater in 1998, 2001-2002, 2004 and 2007-2012. The last extrusion of andesitic block-lava is the subject of our study.

2. VIDEO MONITORING SYSTEM (VMS) OF VOLCAN DE COLIMA AND THE METHODOLOGY OF STUDY OF THE LAVA DOME GROWTH

The presence of a new lava dome in the central part of the crater (Figure 2A) was discovered during an airplane flight in early February 2007. The height of the dome was just a few meters and it was covered by a large amount of grey ash. Lava dome growth was slow; the extruding lava took four years to fill the crater (Figure 2B, 2C). Monitoring of the lava dome growth, carried out with the VMS installed around the volcano, allowed us to obtain the quantitative characteristics of this process.

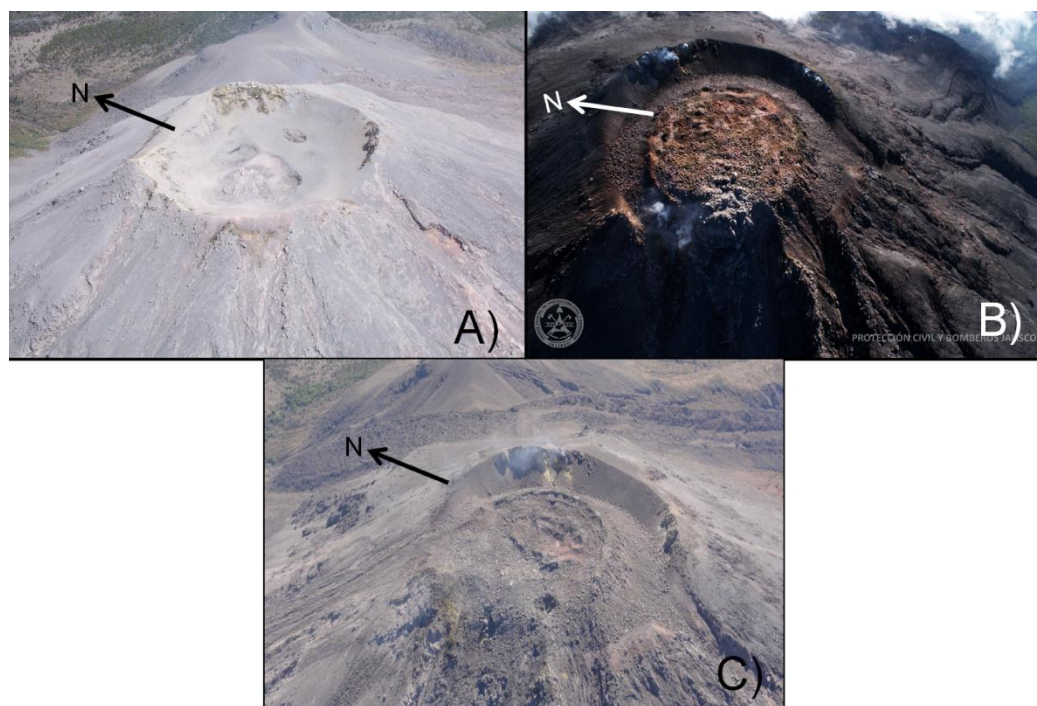


Figure 2. Photographs of the initial stage of lava dome growth (A, February 2007), the intermediate stage (B, October 2010) and the present stage of the crater (C, April 2012). Photographs (A-B) were taken by the Jalisco Civil Protection and (C) was taken by Mauricio Bretón.

2.1. VMS and its Use in the Reconstruction of Eruptive Process

The first two video stations of VMS were installed around the volcano in November 1998 (Nevado and Naranjal). Now the network consists of eight digital video cameras, installed at five monitoring stations. Figure 3 shows the position of video network.

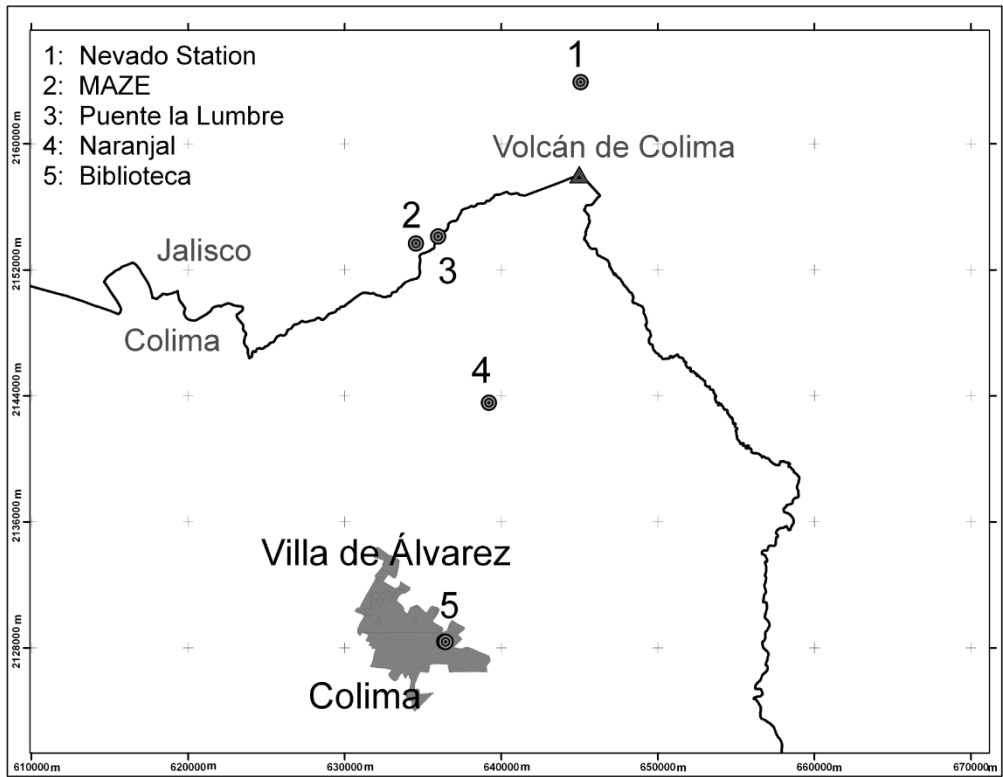


Figure 3. Location of video stations of VMS. The stations are shown with circles.

The video images were transmitted by radio link in real time to the monitoring center situated in the Colima Volcano Observatory in Colima city. Since 2007, video cameras have been synchronized to the world atomic time survey. Table 1 gives the main characteristics of the network.

During the 2007-2012 lava dome growth, 5 video stations were in operations. All stations were equipped with two *FreeWave* radio wireless data transceivers with operating frequency 902-928 MHz, allowing them to transmit images continuously 24 hrs a day. The video stations have different functions. VMS from Nevado station allows reconstructing the temporal variations in the lava dome growth along the N-W profile. El Naranjal station shows the activity that occurs on the southern, southeastern and southwestern sides of the volcano. The MAZE station registers the activity occurring in the south and southwest of the volcano. The Biblioteca station shows an overview of the volcano on the southern side and enables estimation of explosive column heights, and allows monitoring of volcanic plumes. Finally, Puente La Lumbre station, allows us to record the events that occur on the southwestern flank of the volcano.

Table 1. Characteristics of the VMS network

Name of camera	Station	Period of observations	Resolution pixels	Model of camera	Zone of monitoring	Focal length (mm)	Geographic Coordinates	Distance from camera to the crater of volcano
Nevado SONY	Nevado	1998-2010	320x240	Sony Hi8 Handycam Vision CCD-TRV68 NTSC	Crater and dome	3.6-72	19.564° -103.617°	5.3 km
Nevado 221	Nevado	2008-2012	640x480	Axis 221	Volcanic Edifice	3-8	19.564° -103.617°	5.3 km
Nevado 213	Nevado	2006-2012	704x479	Axis 213 PTZ	Crater and Dome	3.5-91	19.564° -103.617°	5.3 km
Naranjal	Naranjal	1998-2012	704x480	Axis 213 PTZ	Lahars, landslides, and volcanic edifice	3.5-91	19.381° -103.674°	15.67 km
MAZE	MAZE	2011-2012	704x480	Axis 213 PTZ	Lahars, landslides, and volcanic edifice	3.5-91	19.473° -103.717°	11.3 km
Puente la Lumbre	Puente La Lumbre	2006-2012	321x240	Axis 221	Lahars in the Río La Lumbre	3-8	19.477° -103.704°	9.9 km
Biblioteca de Ciencias	Biblioteca	2006-2012	640x480	Axis 221	Volcanic edifice and direction of ashes	3-8	19.244° -103.701°	40 km

The station Puente La Lumbre is discarded completely for this job since its purpose is to capture lahars mainly in time of rains. Biblioteca station is also discarded by the distance which lies of the volcano, since its purpose is to detect the direction of flow of ash. MAZE and Naranjal stations are approximately 11 and 15 km from the top of the volcano, however, it is not possible to appreciate the details of the lava dome.

The Nevado station, located closer to the crater (5.7 km), was selected for the reconstruction of the 2007-2012 lava dome growth. This station has had three cameras: Sony, which was in service from 1998 to 2010; the Nevado 221, operating from 2008 to date, and Nevado 213, that is in use from 2006 to date. Each of these cameras recorded approximately 1 image every 3-15 sec depending on weather conditions.

3. METHODOLOGY OF THE IMAGE PROCESSING

The first step was to select a photograph taken at Nevado 221, which showed the profile of Volcán de Colima with the clear sky, in order to distinguish the contours of the volcanic edifice without the interposition of clouds or other objects such as trees, objects, etc. The image, taken on February 24, 2012 at 00: 30: 48 GMT time and shown in Figure 4, became the basic image that was used to rectify and give metric units. The photograph was taken with a camera with focal length of 6.5 mm and resolution of 640 x 480 pixels, in JPG format, at a

distance of approximately 5700 m (Google Earth, 2010) from to center the crater of the volcano.



Figure 4. Basic image of the volcanic edifice of Volcán de Colima, a view from the Nevado 221.

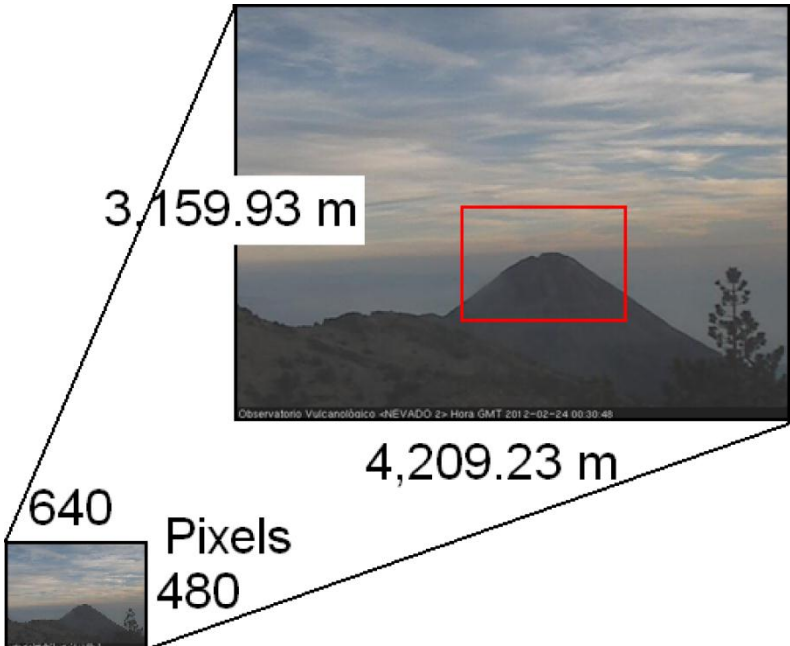


Figure 5. Conversion of pixels to meters of the basic image.

Our supplier *Axis Communications* cameras through their web site, provided a tool to transform the image size from pixels to meters. A photograph of 640 x 480 pixels is 4209.23 m x 3159.93 m. After this operation, the image, corresponding to the upper part of the volcano, had metric units inside the red frame, shown in Figure 5. This image was used as a base to attach the rest of the photos and to draw the outline of the crater and the dome.

4. QUANTITATIVE CHARACTERISTICS OF THE LAVA DOME GROWTH

4.1. Reconstruction of the Lava Dome Growth (E-W Profile)

Photo, taken on October 29, 2007 with resolution of 3008 x 2000 pixels, was overlapped to basic image (Figure 5) using the ArcGis 9.2 Geoprocessing Toolbox. It was used as the base image for the outline of the crater of Volcán de Colima. The video images within this frame were used to reconstruct the dome profiles during the 2007-2012 lava dome growth. Some images were rotated a few degrees to the counter-clockwise with respect to the base image (Figure 6).

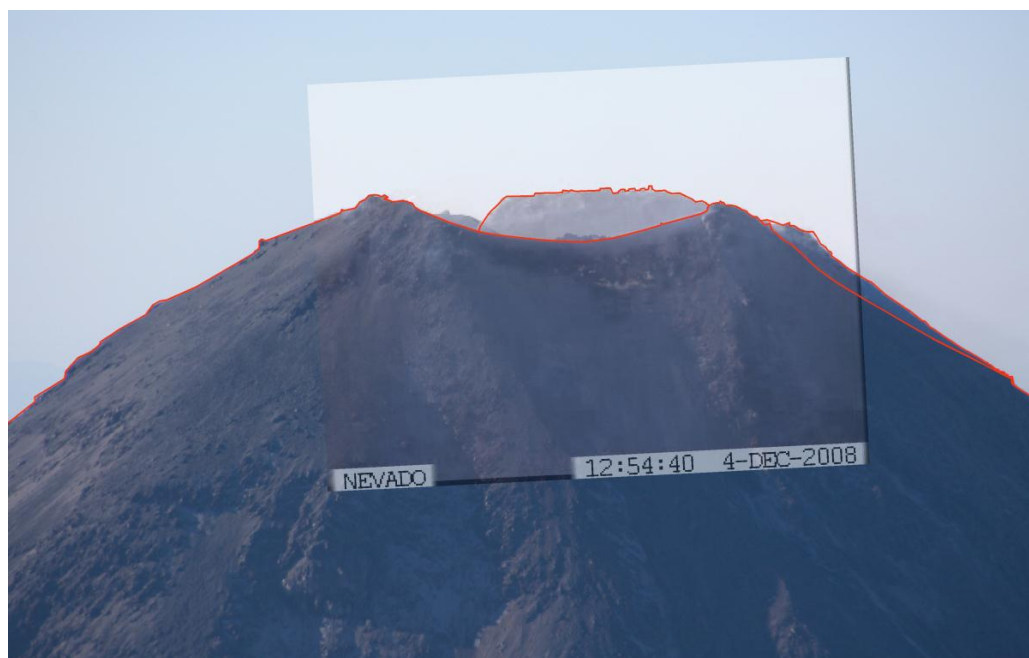


Figure 6. Illustration of the overlap of the base image, taken October 27, 2007, with the image, taken on 4 December, 2008.

The outline of the visible part of the dome was digitized with the help of the program *ArcGis 9.2*, the outlines of the crater were stored as the ArcGIS shapefiles. The images, that were used to digitize the outline of the dome, were taken by the Sony Hi8 fixed camera.

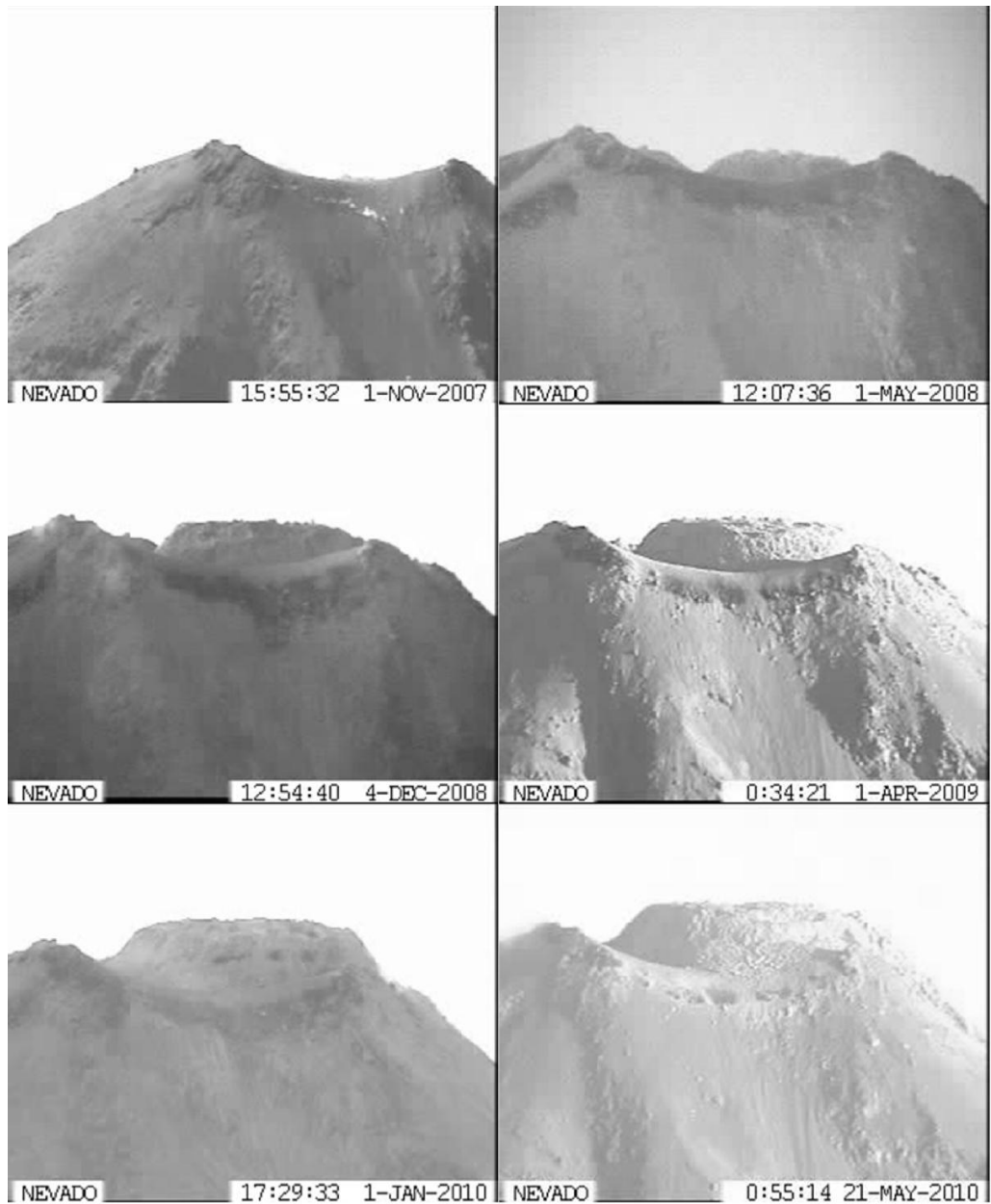


Figure 7. Video images obtained during the 2007-2010 lava dome growth at Nevado Sony station.

A sequence of images, captured by Nevado Sony station between November 2007 and May 2010, is shown in Figure 7; a sequence of images, taken from April 2010 to April 2012 at Nevado 213 station, is shown in Figure 8. In this sequences of images it is possible to appreciate the evolution of the dome growth of the Volcán de Colima.

The final diagram of the contours of the growing lava dome (E-W profile) is shown in Figure 9. It illustrates slightly asymmetrical lava dome growth with more intense growth in the western part. The western side of the crater was completely filled with the dome while the

eastern side of the crater was not filled completely. As a result, the western side of the dome began to collapse in February 2010 generating numerous rockfalls.

The height of the dome was calculated using a reference point (RP) on the rim of the crater estimated from the composed digital model of the volcanic edifice based on the models of INEGI (2010) and LIDAR (2008).

This reference point was chosen at the north border of the crater at the height 3,795 m (a.s.l.) taken with a precision of $\pm 1\text{m}$. Earlier Gonzalez-Mellado (2011) showed that the initial point of the dome growth within the crater had the height 20 m less than our estimation of the reference point on the rim of the crater.

So we corrected the position of our reference point considering it as 3,775 m to have possibility to calculate the heights of the dome from the floor of the crater.

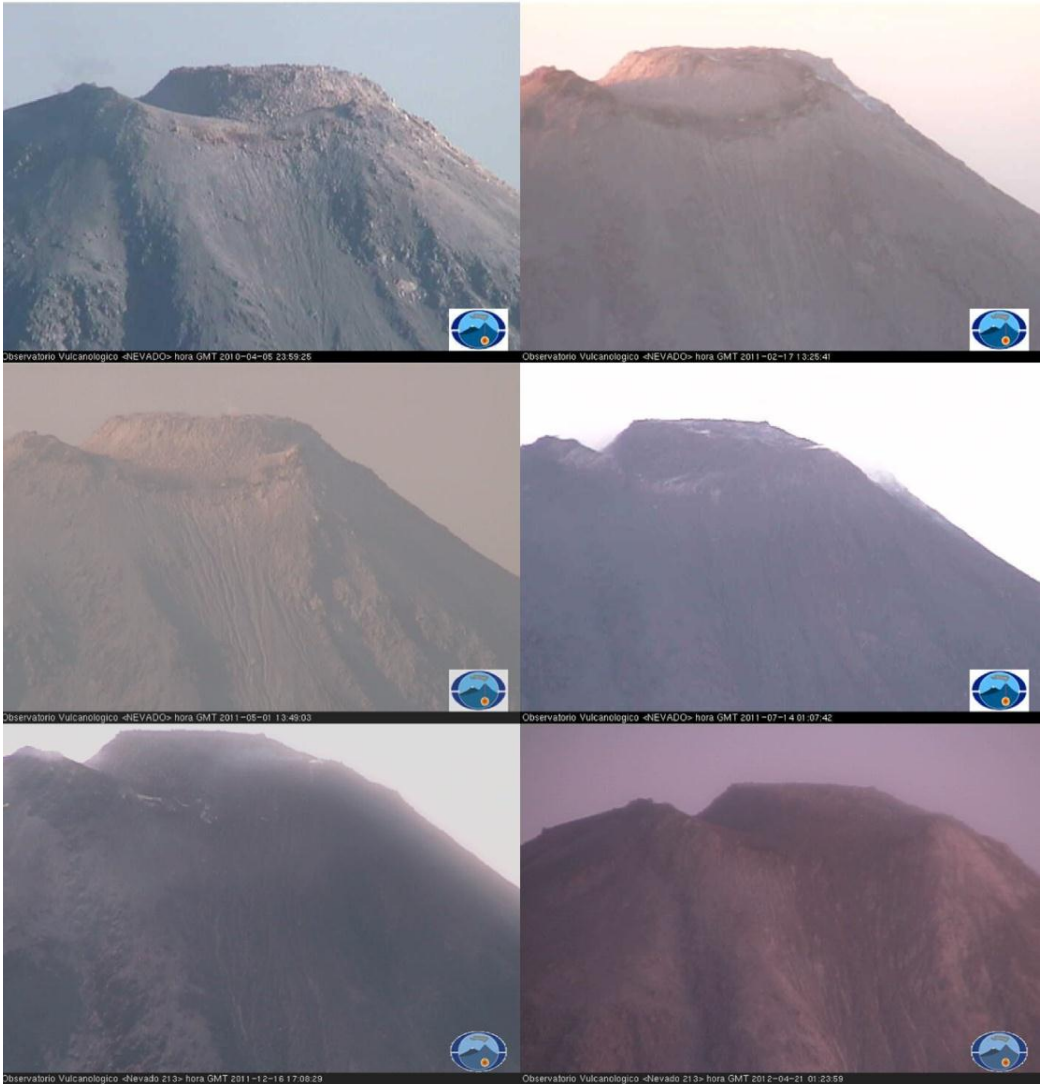


Figure 8. Video images obtained during the 2010-2012 lava dome growth at Nevado 213 station.

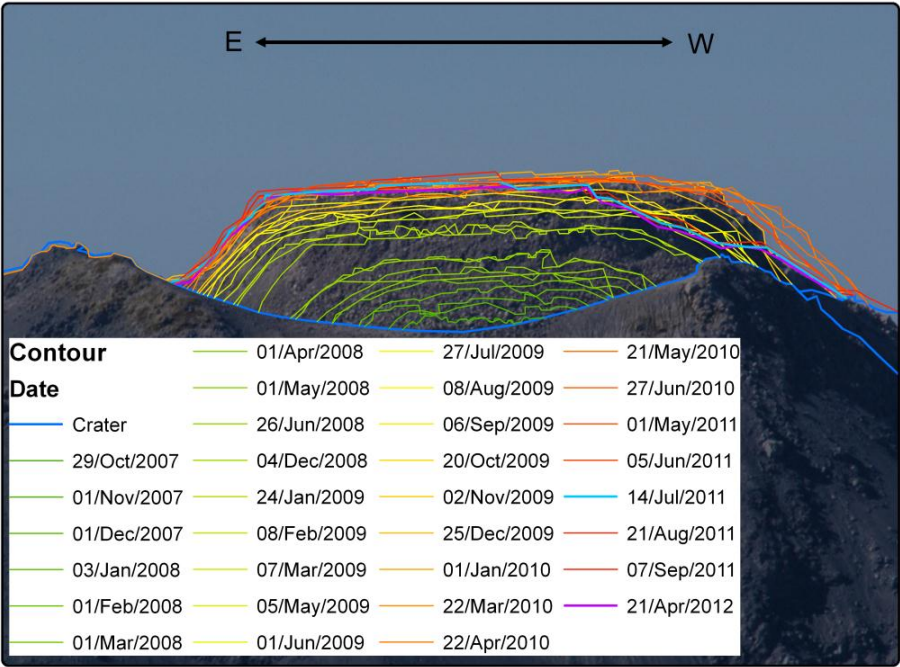


Figure 9. The contours of growing lava dome along the E-W profile between 2007 and 2010. (Nevado station).

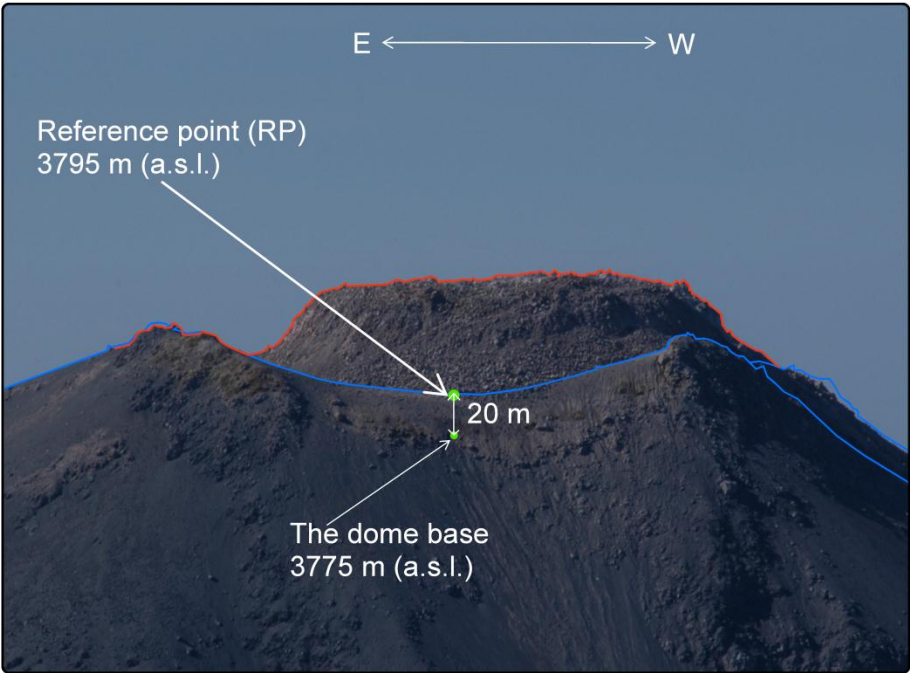


Figure 10. The contours of the crater and the lava dome and the position of the reference point (RP). The contour of the crater is shown with a blue line and the contour of the 2010 dome is shown with a red line.

Figure 11A and Figure 11B show the profile of the rim of the crater based on the composed model and the profile of the 2010 dome obtained from the VMS observation. The absolute heights of the selected points of the dome profile shown in Figure 11B were calculated basing on the position of the reference point.

4.2. Dynamics of the Growth of Lava Dome

The process of the dome growth was especially investigated for two zones of the dome selected in the central and western parts of the dome surface. We call them perspectives A and B (Figure 12).

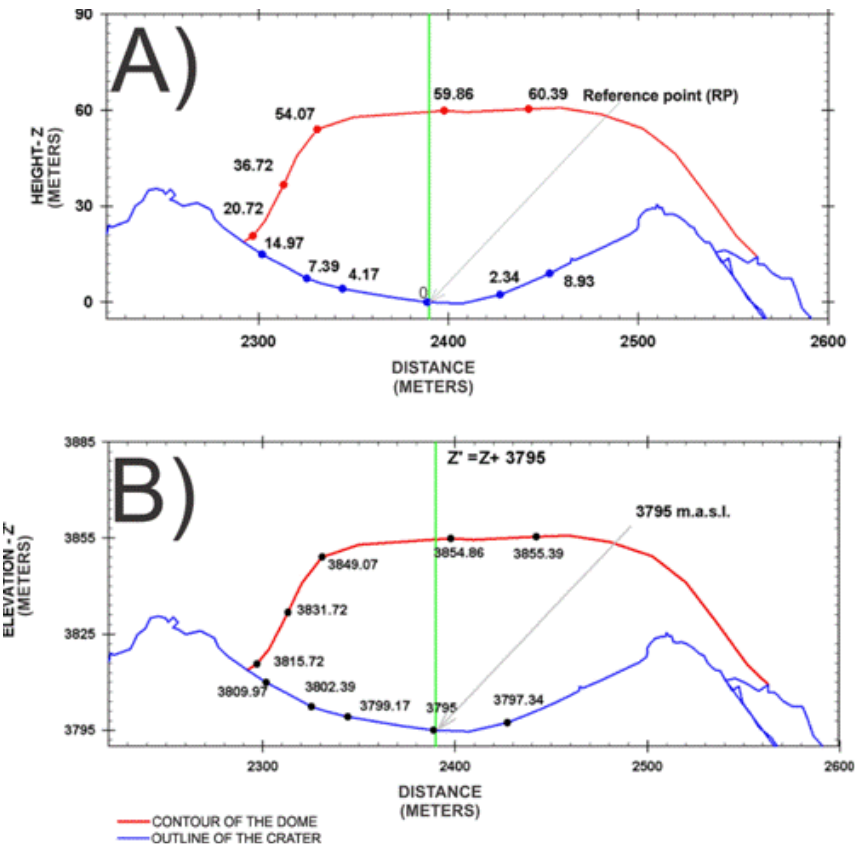


Figure 11. The height calculation for the dome profiles. The indicated points on the surface of the dome were used later for calculation of the volume of the dome.

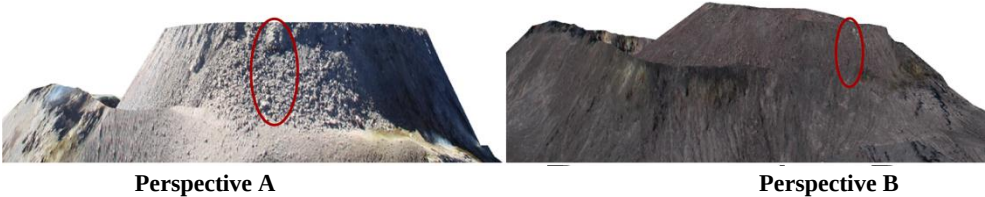


Figure 12. Position of two points on the dome selected for illustration of the lava dome growth.

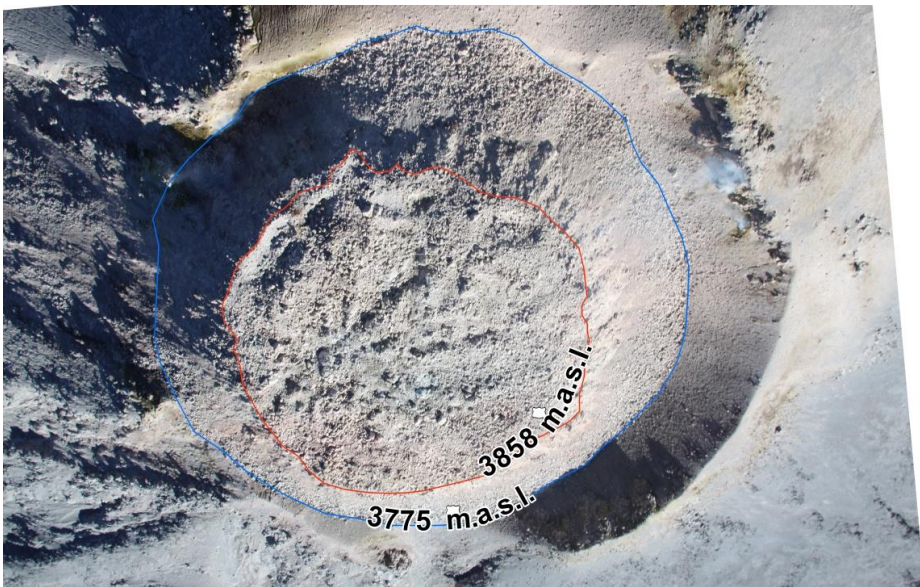


Figure 13. The contours of the base of the crater (blue line) and the dome surface (red line) based on the photograph of the crater zone, taken by the Jalisco Civil Protection on January 2010.

Tables 2 and 3 present the dome height variations during the periods from 29 October 2007 till 21 April 2011 in the central part of the dome (Table 2) and from March 2009 till 21 April 2012 on the western side of the dome (Table 3). The height was calculated from the base of crater shown in Figure 13. Figure 14 and 15 show the variations in the dome growth for these two zones during the period from 2007 to 2012. Figure 14 demonstrates that the growth of the central part of the dome in the crater of Volcán de Colima was continuous till July 2009. In August 2009, the dome began to lose material and its height declined in 2.90 m. Then the growth of the dome was unstable till January 2010, the periods of the dome building were changed with periods of the dome destruction. In January 2010, the dome reached its maximum height of 83.93 ± 3.20 m. In the period between January 2010 up to April 2011, the height of the central region of the dome decreased to 77.40 m demonstrating a reduction of 6.53 ± 0.56 m. On the western side of the dome (Figure 15 and Table 3) a maximum height of 75.85 m was reached in May 2011. After the explosion of 21 June 2011, this side of the dome lost about 16.8 m of its height.

Table 2. Growth of the dome in its central part (Perspective "A")

Date	Elevation (M ASL)	Height of Dome (m)	Growth
			(m)
Nov-07	3801.37	26.37	
Dec-07	3806.29	31.29	4.92
Jan-08	3809.77	34.77	3.48
Feb-08	3812.84	37.84	3.07
Mar-08	3815.31	40.31	2.47
Apr-08	3819.57	44.57	4.26
May-08	3820.44	45.44	0.87
Jun-08	3827.31	52.31	6.87

Date	Elevation (M ASL)	Height of Dome (m)	Growth (m)
Dec-08	3837.08	62.08	9.77
Jan-09	3837.6	62.6	0.52
Feb-09	3838.41	63.41	0.81
Mar-09	3843.15	68.15	4.74
May-09	3844	69.00	0.85
Jun-09	3844	69.00	0.00
Jul-09	3851.25	76.25	7.25
Aug-09	3848.35	73.35	-2.9
Sep-09	3849.81	74.81	1.46
Oct-09	3851.25	76.25	1.44
Nov-09	3854.47	79.47	3.22
Dec-09	3856.88	81.88	2.41
Jan-10	3858.93	83.93	2.05
Mar-10	3857.16	82.16	-1.77
Apr-10	3854.9	79.90	-2.26
May-10	3855.25	80.25	0.35
Jun-10	3855.39	80.39	0.14
May-11	3856.78	81.78	1.39
Jun-11	3855.75	80.75	-1.03
Jul-11	3854.54	79.54	-1.21
Aug-11	3858.31	83.31	1.53
Sep-11	3853.87	78.87	-4.44
Abr-12	3852.4	77.40	-1.47

Table 3. Growth of the dome on its western side (Perspective "B")

Date	Elevation (MASL)	Height of Dome (m)	Growth (m)
May-09	3834.37	59.37	8.66
Jun-09	3835.76	60.76	1.39
Jul-09	3830.77	55.77	-4.99
Aug-09	3835.93	60.93	5.16
Sep-09	3830.74	55.74	-5.19
Oct-09	3843.51	68.51	12.77
Nov-09	3841.81	66.81	-1.70
Dec-09	3851.70	76.70	9.89
Jan-10	3851.88	76.88	0.18
Mar-10	3852.10	77.10	0.22
Apr-10	3851.13	76.13	-0.97
May-10	3850.80	75.80	-0.33
Jun-10	3848.81	73.81	-1.99
May-11	3850.03	75.03	1.22
Jun-11	3846.85	71.85	-3.18
Jul-11	3830.05	55.05	-16.80
Aug-11	3830.20	55.20	0.15
Sep-11	3830.21	55.21	0.01
Apr-12	3828.89	53.89	-1.32

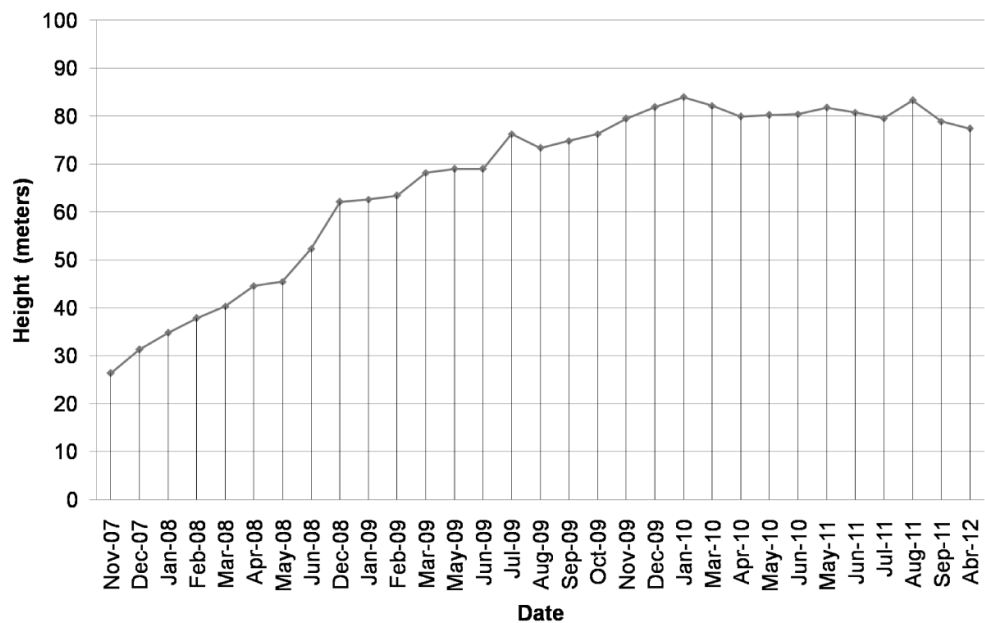


Figure 14. The growth of the central part of the lava dome between November 2007 and April 2012.

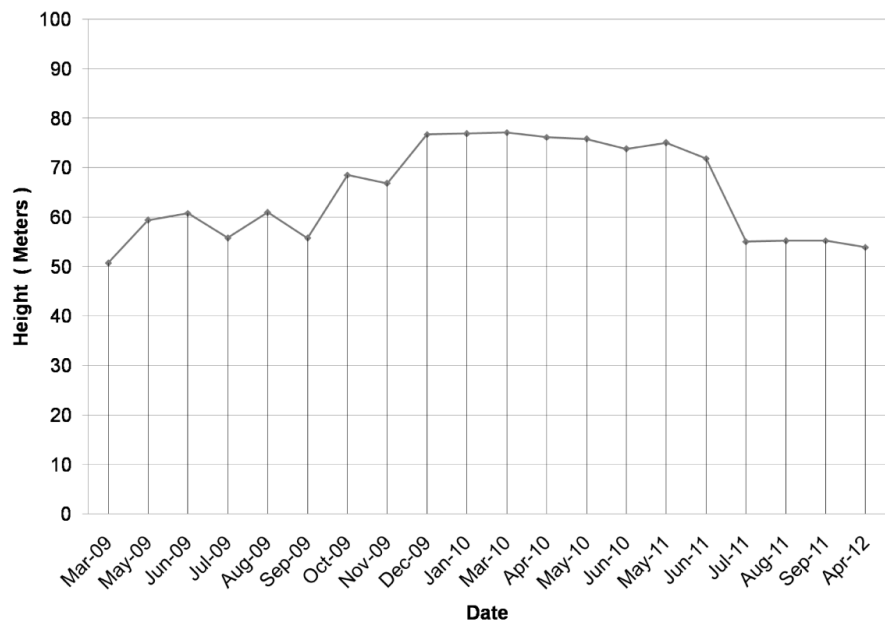


Figure 15. The growth of the western part of the lava dome from March 2009 till April 2012.

The mean rates of the dome growth for two zones of the dome (perspectives A and B, see Figure 12) were calculated for the period from May 2009 until January 2010 when the dome reached its maximum height. The results showed that the rate of dome growth in the central part of the dome (perspective A) was equal to 0.22 m/month and on the western side

(perspective B), the rate was of 0.30 m/month. The mean rate of the dome building from May 2009 to January 2010 may be considered as 0.26 m/month.

4.3. Calculation of Area and Volume of the Dome

The calculation of the areas and the volumes of the dome was performed using a technique defined by Andres de Pablo et al (2011) which requires georeferencing of the images and digitalizing them utilizing the ArcGis program. A photo taken from helicopter at an elevation superior to the 4300 m (a.s.l.) in January 2010 (Figure 13) was used to calculate the volume of the dome of the Colima Volcano. Two concentric curves, one corresponding to the superior base of the dome with an elevation of 3858 m (a.s.l.) and the other corresponding to the inferior base of the dome, located at the elevation of 3775 m (a.s.l.) were digitized (Figure 13).

The calculation of the area and volume of the dome was done considering the dome as a cone limited with two circles corresponding to the curves of the superior and inferior bases of the dome shown in Figure 13. The radii of these circles were estimated equal to 113 ± 11.52 m (inferior base) and 88.5 ± 9.48 m (superior base). It allowed to estimate the area of the bases of the dome equal to $4.01 \pm 0.008 \times 10^4 \text{ m}^2$ (inferior base) and $2.46 \pm 0.008 \times 10^4 \text{ m}^2$ (superior base).

The volume of the dome was calculated with the help of the SIGMAPLOT program which allowed considering a 3D model of the dome (Figure 16).

Assuming the dome as a roughly circular truncated cone, we obtained an area of $1.19 \times 10^6 \text{ m}^2$, and a volume of $2.69 \times 10^6 \text{ m}^3$. The errors in the estimation of the area σ_A and volume σ_V were calculated using equation (1) and (2):

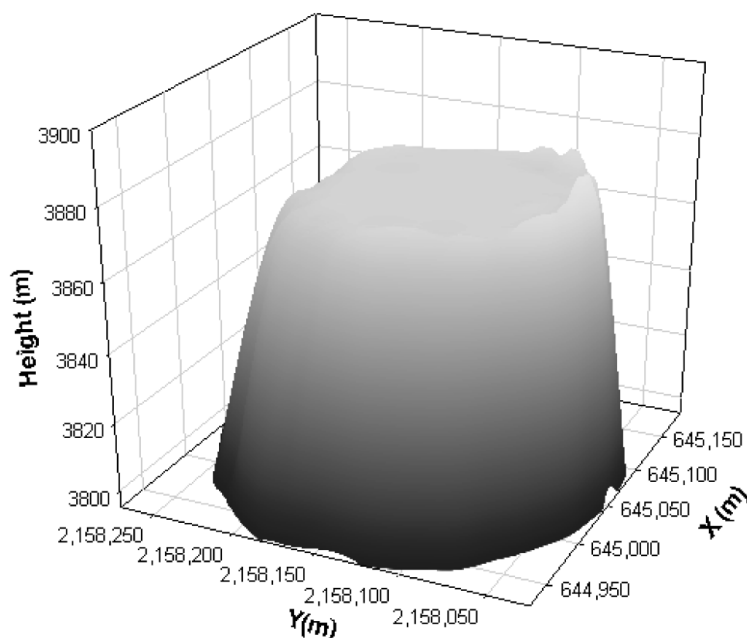


Figure 16. Model of the 2010 lava dome of Volcán de Colima plotting with SIGMAPLOT.

$$\sigma_A^2 = \left(\frac{\partial A}{\partial R}\right)^2 \sigma_R^2 + \left(\frac{\partial A}{\partial r}\right)^2 \sigma_r^2 + \left(\frac{\partial A}{\partial h}\right)^2 \sigma_h^2 \quad (1)$$

$$\sigma_V^2 = \left(\frac{\partial V}{\partial R}\right)^2 \sigma_R^2 + \left(\frac{\partial V}{\partial r}\right)^2 \sigma_r^2 + \left(\frac{\partial V}{\partial h}\right)^2 \sigma_h^2 \quad (2)$$

where σ_R , σ_r , and σ_h are standard error of upper radius, lower radius and height respectively. We obtained $\sigma_A = 0.00447 \times 10^5 \text{ m}^2$ and $\sigma_V = 0.6348 \times 10^6 \text{ m}^3$. Finally, the volume was estimated equal to $2.69 \pm 0.6348 \times 10^6 \text{ m}^3$ and the area, equal to $1.19 \pm 0.00447 \times 10^5 \text{ m}^2$.

CONCLUSION

The observations carried out by the VMS around Volcán de Colima during 2007-2012 and the 3D modeling of the 2010 dome allowed us to obtain the following results:

The mean rate of the dome building from May 2009 to January 2010, when it reached its maximum height, was 0.26 m/month.

The volume of the 2010 lava dome was estimated equal to $2.69 \pm 0.6348 \times 10^6 \text{ m}^3$ and the area, equal to $1.19 \pm 0.00447 \times 10^5 \text{ m}^2$.

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Chapter 9

SEISMIC SIGNALS ASSOCIATED WITH GEOTHERMAL ACTIVITY WITHIN VOLCANIC ENVIRONMENT

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ABSTRACT

Geothermal system is a natural heat transfer within a confined volume of the Earth's crust where heat is transported from a heat source to a heat sink, usually the free surface. Surface manifestations of this process may be hot springs, geysers, mud volcanoes. These manifestations are associated with a special class of the natural seismic events. They occur as a continuous geothermal tremor (noise) and/or the distinct short seismic events. This review presents the examples of these seismic signals observed at different geothermal systems of the world. Comparing with the seismic events associated with volcanic eruptions, the geothermal seismic events are of higher-frequency indicating the smaller size sources of seismic vibrations.

INTRODUCTION

Volcanic environment is a region that includes the variety of volcanic structures such as active volcanoes, dormant volcanoes, and geothermal systems. The seismic activity associated with the eruptions of active volcanoes or with the re-activation or failed eruptions of dormant volcanoes is widely described (e.g., Zobin, 2012). Activity of geothermal systems within volcanic environments represents the process that may be named as a mild eruption. The seismicity of these mild eruptions is a subject of the recent review.

Geothermal system is a natural heat transfer within a confined volume of the Earth's crust where heat is transported from a heat source to a heat sink, usually the free surface. The transport of heat to the surface in volcanic regions occurs by a convective process associated with the rise of hot magma. The heat is passed from the cooling magma in the chamber to the

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surrounding country rocks. Where these chambers or the rock above the chamber contain groundwater bodies, the water expands and rises buoyantly, to be replaced by cold water flowing in from the sides. This process establishes a geothermal circularly system (Hochstein and Browne, 2000; McCall, 2005).

The majority of surface manifestations of heat discharge in volcanic regions may be classified as diffusive, continuous, or intermittent. Diffusive heat transfer by thermal conduction usually results in elevated near surface temperatures and can be recognized mainly by temperature surveys in shallow boreholes. A mud volcano is one of the special surface manifestations in the regime of diffusive heat discharge. The most common manifestations of continuous heat discharge are through convection in hot water springs. Another example is a hot steam at the saturation temperature through fumaroles. Geysers are spectacular manifestations discharging intermittently a mixture of boiling water, gas, and steam (Hochstein and Browne, 2000). Discharge of hot water and steam to the surface may produce the seismic signals.

In this Chapter I present a review of the seismic signals associated with the natural manifestations of geothermal activity within the volcanic environment. The natural or induced seismicity associated with the manifestations of geothermal activity within non-volcanic regions is not discussed.

POSITION OF GEOTHERMAL SYSTEMS WITHIN VOLCANIC ENVIRONMENT

Figure 1 shows an example of the position of geothermal systems within volcanic environment. Kamchatka Peninsula is situated at the northern end of the Kurile-Kamchatka island arc and is characterized by a high level of the seismic and volcanic activity as the result of interaction between the continental North American plate and the subducting oceanic Pacific plate.

The volcanic environment of Kamchatka is separated into two volcanic belts: Eastern (E_K) and Central (C_K). The Central belt is represented by only one active volcano but a great number of dormant volcanoes and hot springs, and two high-temperature geothermal systems. The Eastern belt is more active and consists of 28 active volcanoes, 9 high-temperature geothermal systems (one of them is the famous Geyser valley), and the numerous dormant volcanoes and hot springs. Both volcanic belts were formed during the Quaternary period (Leonov, 2001).

Figure 1 shows that all high-temperature geothermal systems in Kamchatka are situated along the chains of active volcanoes. Hot fumarolic fields with temperatures higher than 400°C are acting during centuries in the craters of Tolbachik, Mutnovsky, Avachinsky, Zhupanovsky and Gorely. Acid and hot lakes are located in craters of Maly Semiachik, Mutnovsky and Gorely volcanoes. Temperatures as high as 220°C and 240° C have been measured in strong flank fumaroles of Koryaksky and Kizimen volcanoes. Boiling-point fumarolic fields are located at volcanoes of the Eastern volcanic belt such as Komarov, Bolshoy Semyachik, Burlyaschii, Kikhpinych, Dzenzur, Zheltovsky, Ksudach, Kambalny and Koshelevsky. A fumarolic field exists at the slope of Ichinsky volcano, the only active volcano in the Central volcanic belt.

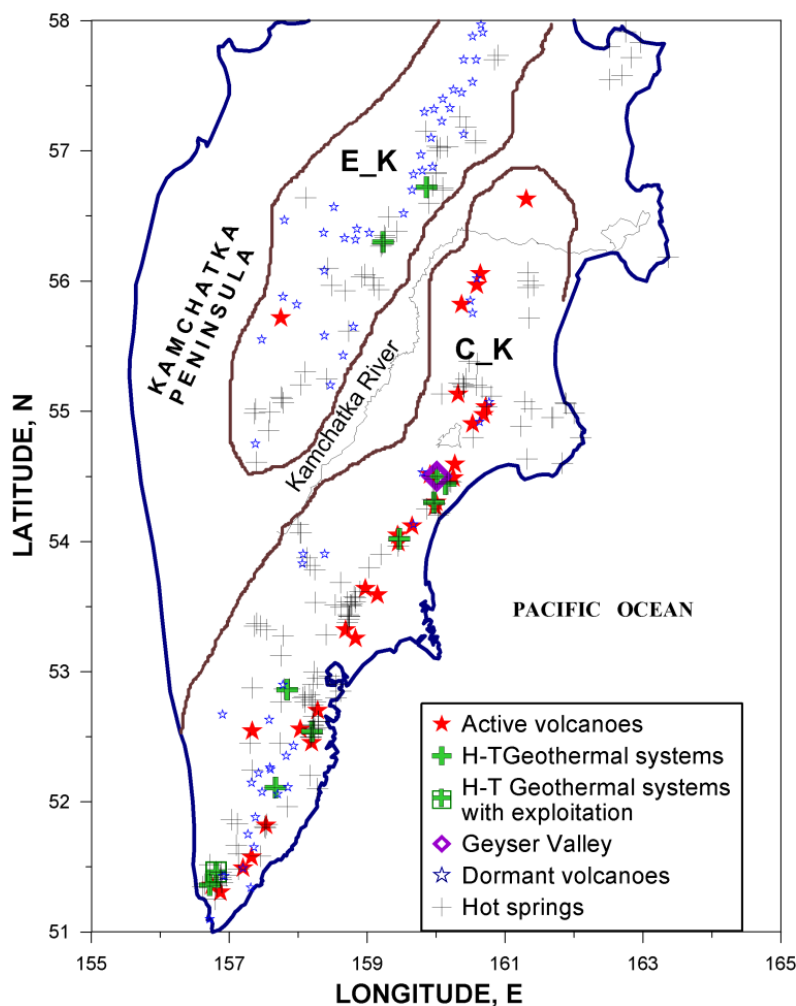


Figure 1. Volcanic environment of Kamchatka Peninsula, Russia. There are shown active and dormant volcanoes, hot springs and high-temperature (H-T) geothermal systems, and Geyser Valley. E_K, Eastern Volcanic Belt; C_K, Central Volcanic Belt. Catalog of active and dormant volcanoes is taken from Simkin and Siebert (1994). Catalog of hot springs is taken from Kiryukhin et al. (2010). Catalog of high-temperature geothermal systems is taken from (Sugrobov and Yankovsky, 1991).

Four high-temperature systems situated within the Eastern volcanic belt have been drilled, and the first geothermal plant in Pauzhetka in southern Kamchatka has been in operation since the end of 1966. Mutnovsky geothermal field provides now >60 MW of electric power (Taran, 2009).

STRUCTURE OF GEOTHERMAL SYSTEM

Figure 2 shows two conceptual models of the structure of geothermal systems (Hochstein and Browne, 2000). The general sketch of the deep structure of geothermal system (Figure 2A) displays the main elements of the system. Heat transfer from the magma reservoir through conductively heated zone is realized by convective processes. The convecting heat

transfer medium is dominantly infiltrated surface water. The fluids that transfer most heat are magma and magmatic gases. The surface manifestations are spatially zoned with respect to their volcanic center and are represented by solfataras, fumaroles, hot acid lakes, acid hot springs, and acid streams. At lower elevations, minor thermal springs may discharge neutral pH, chloride, or sometimes bicarbonate-chloride waters.

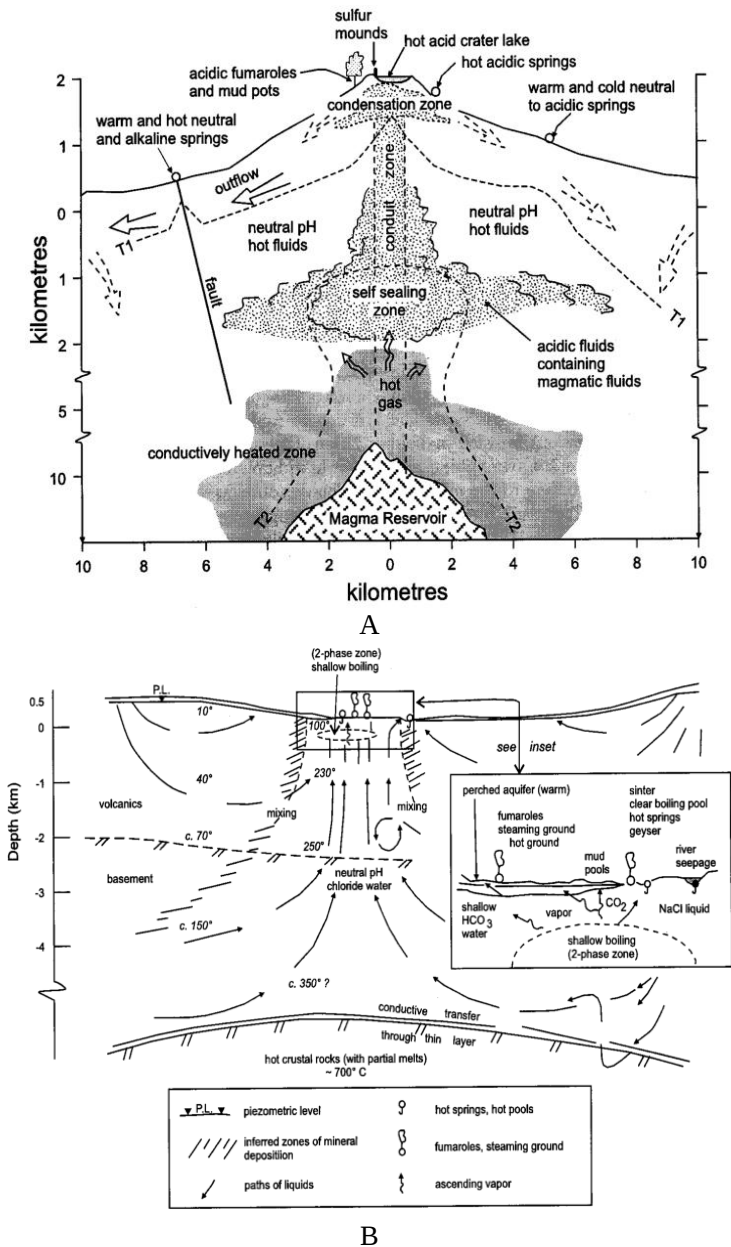


Figure 2. Conceptual models of the deep structure of geothermal systems and their surface manifestations. A, a general sketch. Broken lines with T1 and T2 represent inferred isotherms for 150°C and 350°C, respectively. B, conceptual model of a liquid dominated geothermal system. After Hochstein and Browne (2000).

The special case of a liquid dominated geothermal system is shown in Figure 2B. The heat source is an extensive layer of hot crustal rocks that contains some partial melts and host intrusions. Because of their high recharge rate, the hot fluids can ascend close to or reach the surface. This thermal regime favors geothermal eruptions. The surface manifestations of this type of system are boiling hot springs and geysers.

NATURAL SEISMICITY ASSOCIATED WITH GEOTHERMAL SYSTEMS

Diffusive Heat Discharge (Mud Volcano)

As was mentioned above, the mud volcanoes represent the surface manifestations in the regions of diffusive heat discharge. The nature of sedimentary mud volcanism is supposed to be driven by gravitational sediment imbalance and fluid overpressures (Kopf, 2002). The only case of the birth of a mud volcano recorded seismically, was Lusi mud volcano.

The 29th of May 2006 at around 05 h (28 May, 22 h GMT, local time – 7 h) the Sidoarjo district in East Java, Indonesia witnessed the sudden birth of numerous gas and mud vents on the ground. These eruption sites extended over a distance of more than a kilometer forming a NE–SW alignment. The eruptive bursts of hot water and steams were dramatic with a distinct geyser-like cycle of active and passive periods. Within a few days a prominent crater formed and the initial vents were quickly covered by the large volumes of erupted boiling mud. The flow-rates from the main crater reached 180 000 m³/day. The largest crater of new mud volcano was named Lusi [acronym from LUMPur (mud) and SIdoarjo (the district name)] (Mazzini et al., 2007; 2012; Savolo et al., 2009).

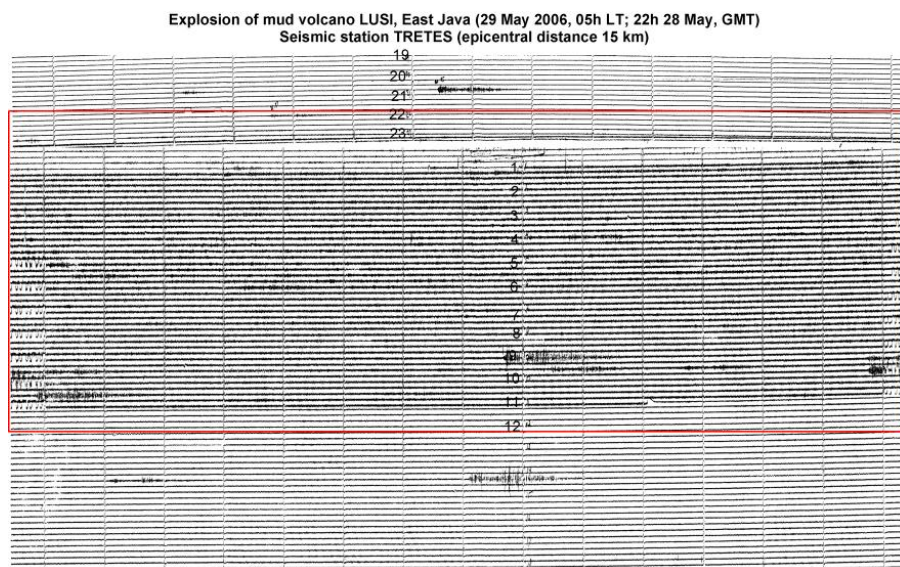


Figure 3. Short-period seismic record, associated with the eruption of LUSI mud volcano, Indonesia, obtained during 28 and 29 May 2006 at the seismic station Tretes situated at a distance of 15 km from the mud volcano. The 10-h interval from 1 to 11 h 29 May (GMT) characterized by high-frequency low-amplitude continuous seismic signals, recorded in about 3 hrs after the beginning of the eruption, is shown in rectangle. Courtesy of Dr. Bambang P. Istadi, Energi Mega Persada.

The samplings of the fluids and the solids erupted by Lusi showed the presence of mud breccia and significant amounts of methane, typical of mud volcanoes worldwide. Preliminary analyses of erupted material indicated that the main source of clay and water were the overpressured units located at ~1500–1800 m that were rapidly buried and under-compacted (Mazzini et al., 2007).

Figure 3 shows the seismic record obtained during 28 and 29 May at Tretes BMG station situated at a distance of 15 km from the mud volcano. It is clear seen the 10-h interval from 1 to 11 h 29 May (GMT) characterized by high-frequency low-amplitude continuous seismic signals recorded in about 3 hrs after the beginning of the eruption. A few high-frequency high-amplitude signals of duration up to 2 min were recorded during this period and later. The largest of them were recorded at 08:59 and 14:14 of 29 May (GMT).

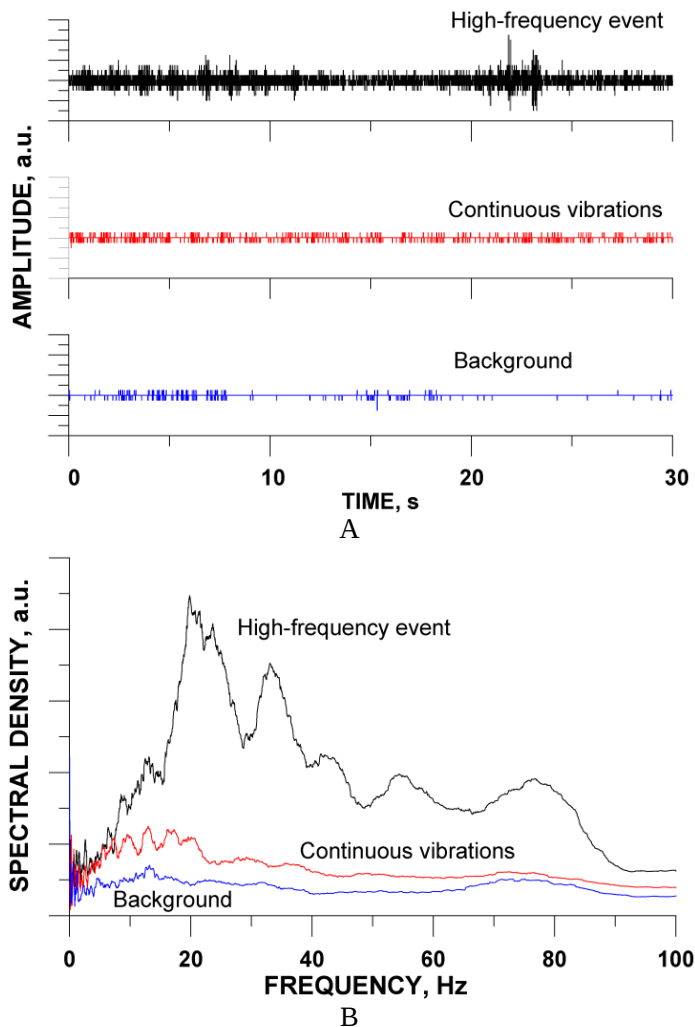


Figure 4. Seismic records (A) and Fourier spectra (B) of 200 Hz sampling data associated with the eruption of LUSI mud volcano, Indonesia on 29 May 2006. Short-period seismic station Tretes, 15 km from the mud volcano. In A, the upper line shows the high-frequency event recorded at 09 h, the middle line shows the continuous vibrations recorded at 10 h, and the lower line shows the background vibrations recorded at 12 h (See Figure 3). Courtesy of Dr. Bambang P. Istadi, Energi Mega Persada.

Fourier spectra, calculated for 30-s time windows of low-amplitude continuous records, high-amplitude events, and of background seismic record (Figure 4A) are shown in Figure 4B. High-frequency events were characterized by the dominant frequency bands between 15–23 Hz and 23–40 Hz. The dominant frequencies of low-amplitude continuous signals were between 5 and 20 Hz. For comparison, is shown the spectrum of the background signal having significantly lower spectral amplitude with a peak frequency at 12 Hz; at higher frequencies, well expressed for the continuous signal, the spectral amplitude gradually decreased.

Continuous Heat Discharge

This type of heat discharge is represented by hot water springs forming geothermal systems. Geothermal systems are zones where significant heat and mass redistribution between a magma reservoir and the ground surface takes place (Norton, 1984). Geothermal systems usually appear on active volcanoes in response to the geothermal fluid circulation (i.e. magmatic gases, groundwater, meteoric waters and seawater).

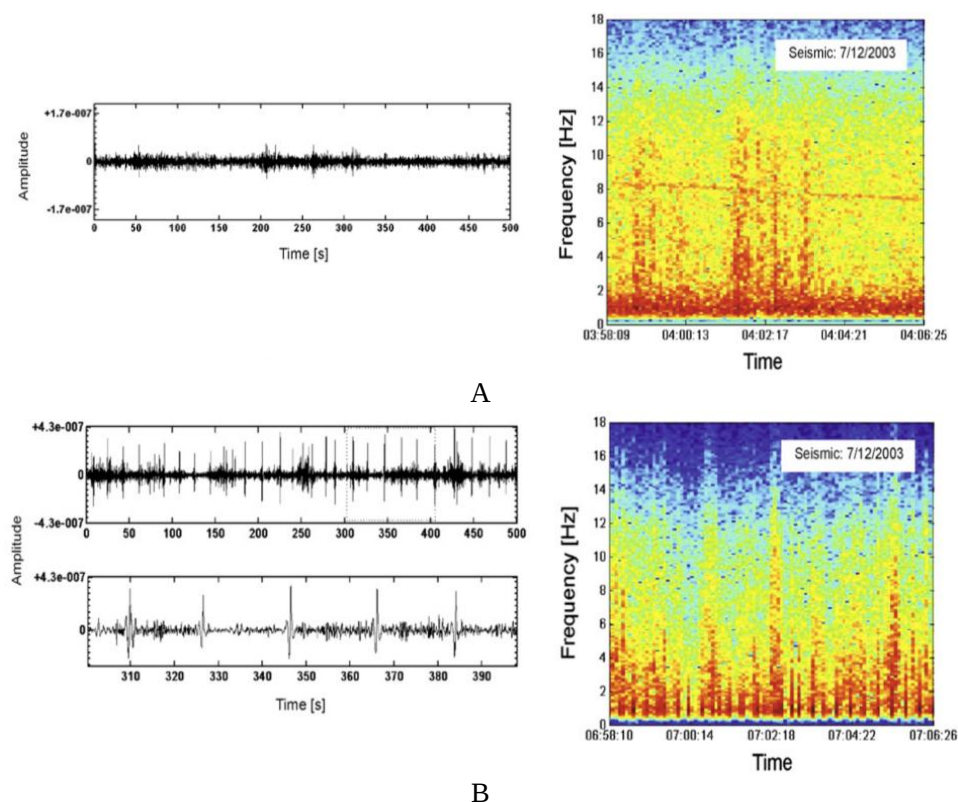


Figure 5. Two types of geothermal tremor (short-period, 100 Hz sampling, seismic signals and corresponding spectrograms) recorded near hornitos of Dallol, Ethiopia, geothermal system on 7 December 2003. A, low-amplitude tremor; B, high-amplitude tremor with strong pulses. In B, the interval between 300 and 400 s is shown also in details. After Carniel et al. (2010).

Dallol, Ethiopia. The Dallol geothermal system is located in the Danakil depression, in the Afar province, Ethiopia. Dallol has a superficial volcanic system associated to posteruptive activity; its last recorded eruption was in 1926. The salt lake, situated within depression, represent the important source of volcanic heat survives, causing the ascent of hot water through the layers of halite and anhydrite. The minerals are dissolved and deposited over the basaltic lava flows, originating fumaroles and formations called hornitos, small rootless spatter cones that often form on the surface of a *basaltic lava* flow (Carniel et al., 2010).

The broadband and short-period seismic instruments that were installed near hornitos of Dallol during December 2003 by Carniel et al. (2010) showed the presence of two regimes of the continuous geothermal tremor with predominant groups of frequencies around 0.5–2 Hz, 2–12 Hz, and 8 Hz (Figure 5). The first regime was characterized by low-amplitude continuous signal (Figure 5A) while the second type of regime was characterized by the presence of repeating strong pulses with duration of about 4.5–5 s, repeated every 2–14 s (Figure 5B).

The continuous data indicate variable shallow processes most likely related with variations in temperature and degassing processes within the shallow geothermal system. The depth of the heat source of perturbations beneath the hornitos was estimated as very superficial, less as 100 m (Carniel et al., 2010).

Uzon, Kamchatka. The Uzon caldera hosts a large geothermal field and numerous hot and cold lakes. Present geothermal activity is concentrated in a 0.2–0.3 x 5 km zone filled with boiling springs, gas-steam jets, mudpots, small mud volcanoes, hot lakes and springs. The most recent phreatic eruption took place in 1989 and created a 14 m wide crater. The Uzon caldera hosts a few Holocene volcanic vents (Leonov et al., 1991).

Figure 6 shows short-period seismic records obtained in the Uzon geothermal system by Kugaenko et al. (2009) during November–December 2007. The geothermal background tremor was characterized by the peak frequency at 3.5 Hz (Figure 6, A and B). There were recorded also 5–7-s-long signals (Figure 6, C and D) characterized by the frequency band of 3–9 Hz. Kugaenko et al. (2009) noted a repetitive nature of these signals.

Satsuma-Iwojima, Japan is a volcanic island located on the northwestern rim of the submarine Kikai Caldera, whose last large eruption occurred about 6300 years ago. Recently, the small ash eruptions were recorded during 1988–2004 (GVP, 2012). The eastern half of the island is occupied by a rhyolitic cone called Iwodake, which has an active crater at its summit of 400 m in diameter.

At the bottom of the crater, there is a gradually expanding fumarolic vent, which continuously and steadily emits high-temperature volcanic gas at nearly 900 °C. In addition to the central vent, there are many small, active fumarolic vents in and around the summit area of Iwodake (Ohminato, 2006).

Broadband seismic observations, carried out during April 1997 by Ohminato (2006), showed the presence of a continuous geothermal tremor (Figure 7). In the periodic seismic signals observed at Satsuma–Iwojima there were observed the frequency peaks and also frequency-dependent amplitude modulation.

Two predominant groups of frequencies: around 6 Hz, and around 20 Hz were identified at ST1 station shown in Figure 7. The 20-Hz peak was common for other 3 broadband stations installed around the crater but the position of the first frequency peak varied from

station to station within 6 to 12 Hz range. Ohminato (2006) considers that the common 20-Hz peaks can be interpreted as a source effect.

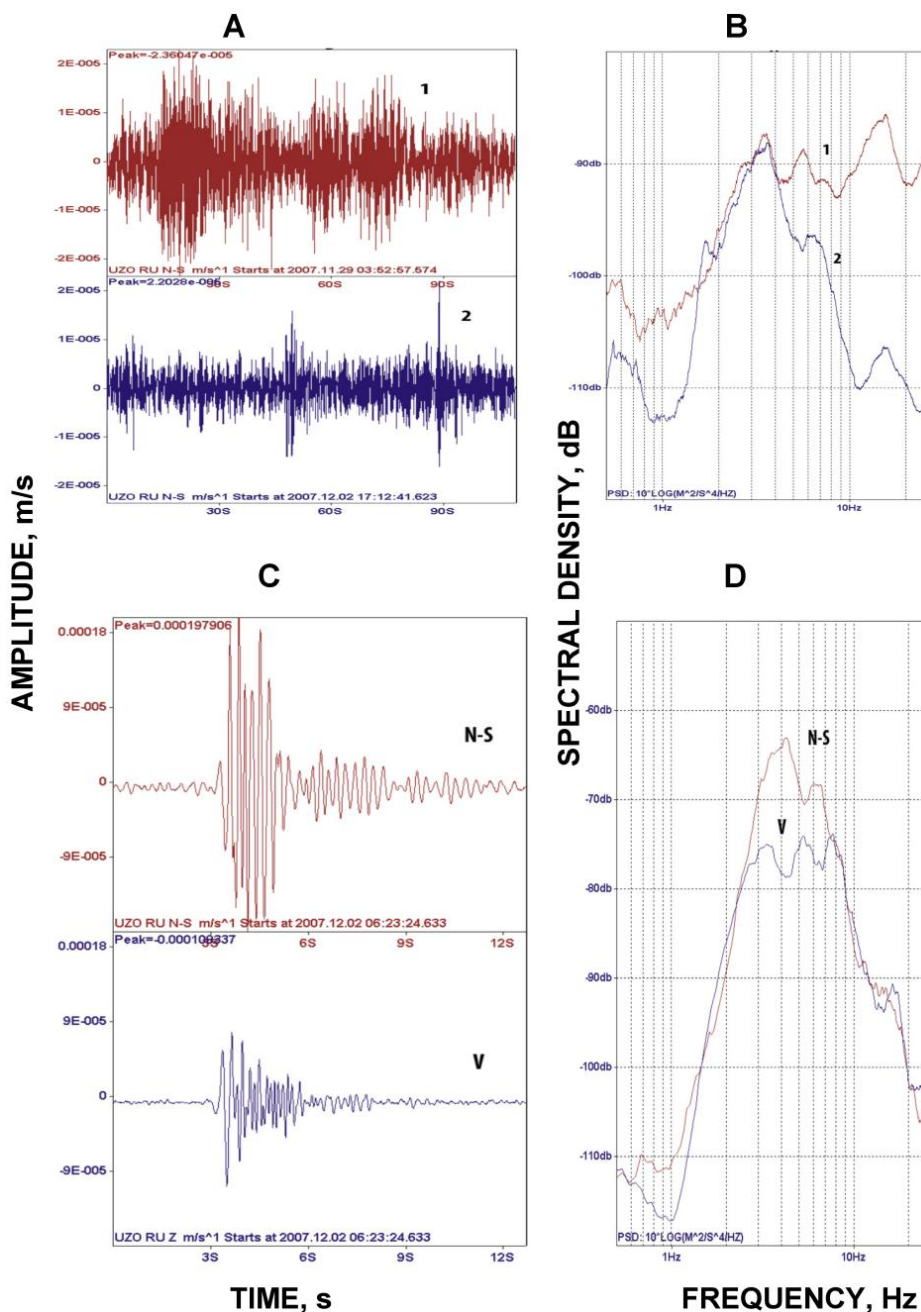


Figure 6. Short-period seismic records and spectrograms of 100 Hz sampling data of geothermal tremor associated with the activity of geothermal system at Caldera Uzon, Russia (seismic station was installed at a distance of about 200 m to the north of the boiling mudpot and hot spring). A and B, geothermal tremors recorded during the windy (1) and calm (2) weather, respectively, and their power spectra; C and D, two-component (N-S, North-South; V, vertical) seismic records of local high-amplitude pulses and their power spectra. After Kugaenko et al. (2009).

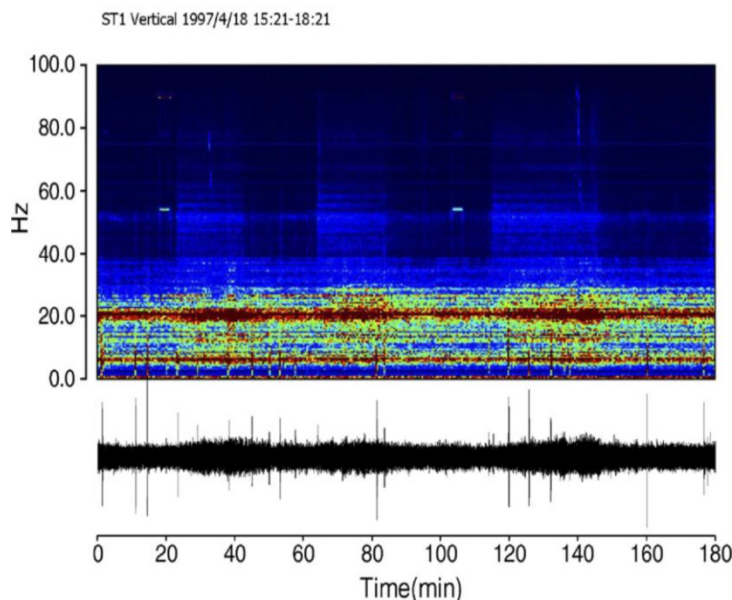


Figure 7. Seismic record (below) and spectrogram (above) of 200 Hz sampling data of geothermal tremor associated with the activity of geothermal system at Satsuma-Iwojima, Japan, volcanic island (broadband seismic station was installed at a distance of about 200 m from the crater). From Ohminato (2006).

Intermittent Heat Discharge

This type of heat discharge is represented by the geyser activity, the phenomena when water and steam spurt from the surface intermittently in a high-reaching fountain. The geyser eruptions are usually accompanied by seismic signals. These seismic signatures vary from one geyser to another. There are two classes of geyser eruptions: fountain and columnar. The fountain eruptions of geysers represent a series of intermittent bursts of bubbles of steam through a surface pool of water. Columnar eruptions of geysers occur as narrow vertical or subvertical columns of ejected water and steam. This type of geysers does not have a surface pool of water (Nicholls and Rinehart, 1967; Kieffer, 1984). The typical seismic signals of the geyser eruptions are described below.

Fountain geyser Strokkur, Iceland. Strokkur's basin is about 12 m in diameter. Its central tube, about 30 cm in diameter, has been built up almost to the surface of the water by deposition of geyserite. Every 10-15 minutes the geyser erupts. When superheated water reaches the surface, the water changes suddenly to steam, and the explosion hurls a small quantity of water to a height of 20 m or more. The eruption is short and continues for a few seconds (Rinehart, 1968). Figure 8 shows the seismic signals recorded a few seconds after the beginning of an eruption of the geyser. Strokkur's seismicity is characterized by a series of short, 1-2 s duration, bursts of oscillations that followed the eruption and are separated by 10-20 s. The successive seismic bursts are not identical. The first bursts in the series are audible at the surface (Rinehart, 1968). Rinehart (1968) considers that the origin of these seismic signals is supposedly associated with sudden refilling of subterranean cavities, which produces a water hammer effect.

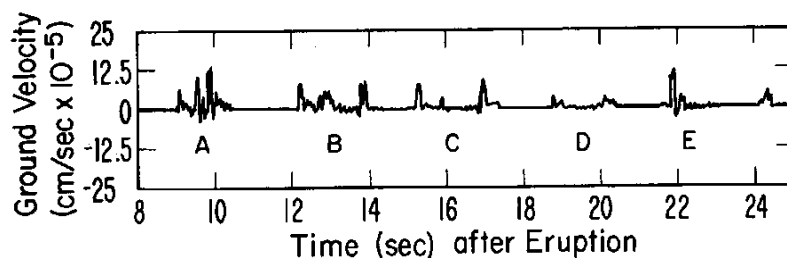


Figure 8. Short-period seismic record of the seismic signals following the 7 July 1967 eruption of Strokkur geyser, Iceland. The seismic station was installed at the rim of geyser. From Rinehart (1968).

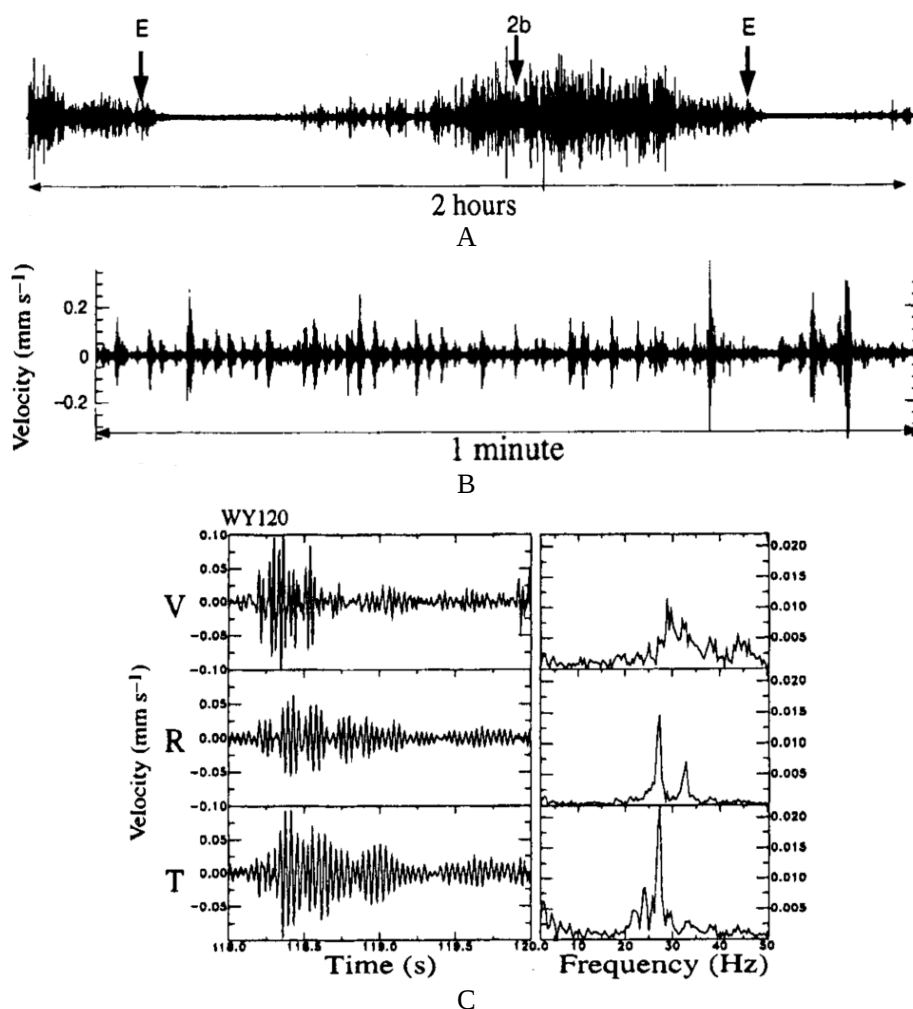


Figure 9. Broadband seismic records associated with the 1992 eruptions of Old Faithful geyser, U.S.A. A, the 2-hrs sequence of the seismic signals associated with two explosions (marked by E) of the geyser and recorded at a distance of about 40 m from the crater; B, the detailed 1-min seismic record of the seismic signals of a sequence of single events preceding the eruption of geyser (this time interval is marked by 2b in A); C, three-component seismic record (left) and corresponding Fourier spectra (right) of a single event similar to shown in B and recorded at a distance of about 30 m from the crater of geyser. After Kedar et al. (1996).

Columnar geyser Old Faithful, Yellowstone. The surface expression of this geyser is a 4-m-high, 60-m wide mound with an opening roughly $2 \times 1 \text{ m}^2$ at the top. The geyser's eruptions are 2-5 min long with the interval between them ranging from 30 to 100 min.

Figure 9A shows broadband seismic record of a sequence of two explosions of the geyser occurring during 2-hour interval. The seismic signals, preceding the geyser eruption, begin about 70 min prior to an eruption and may be divided into three parts: initial low-amplitude vibrations, high-amplitude vibrations, and preliminary low-amplitude vibrations beginning about 12-15 min just prior to the geyser eruption. The eruptions (shown by arrows with E in Figure 9A), were followed by short seismic signals of about 2 min duration. Figure 9B gives a detailed 1-min sequence of the preceding seismic activity. It is seen that this sequence consists of numerous distinct events. Figure 9C shows the 3-component records of one of these distinct events together with corresponding Fourier spectra. The frequency range of the seismic pulses lies between 25-35 Hz (Kedar et al., 1996).

An obvious source of the seismic signals associated with Old Faithful geyser eruptions is supposed to be the repeated collapse of transient steam bubbles in the water column (Kieffer, 1984; Kedar et al., 1998). Kieffer (1984) considers that no evidence that rock fracturing is important during eruption cycles.

DISCUSSION

This short review shows that the geothermal systems situated within volcanic environments are characterized by natural seismic activity. These seismic signals may be described as a continuous low-amplitude geothermal tremor and the distinct high-amplitude seismic events. The origin of these seismic signals is supposed to be related to the movement of hot steam and water bubbles within the surface heated soil or (and) in the water and steam columns.

These signals recorded during so-called mild eruptions may be compared with the seismic signals recorded during volcanic eruptions. Figure 10 shows the variety of the seismic signals that were recorded during the 1998-2011 effusive-explosive eruption of andesitic Volcán de Colima, México, and corresponding Fourier spectra. These volcanic earthquakes were associated with explosive and effusive activity of the volcano. Kieffer (1984) and Kedar et al. (1998) pointed out the similarity between the seismic signals associated with the geyser and volcanic activity.

Kieffer (1984) had compared the seismograms of the banded tremor recorded during the eruption at Karkar volcano, Papua New Guinea, and the seismic records of activity at Old Faithful geyser, Yellowstone National Park and concluded that short monochromatic high-frequency seismic events recorded prior to geyser eruption resemble harmonic volcanic tremor. Kedar et al. (1996) showed that the seismic activity between eruptions at Old Faithful Geyser is composed of a superposition of discrete events originated by pressure pulses inside Old Faithful, in analogy to volcanic long-period events, which are considered the building blocks of continuous volcanic tremor.

Figures 11 and 12 compare the seismic records and spectra of geothermal tremor and distinct events recorded at the Uzon geothermal system, Kamchatka (Kugaenko et al., 2009) with the seismic records and spectra of volcanic tremor (TR) and micro-earthquakes (ME)

which often are defined as long-period events (Varley et al., 2010). It is seen that the seismic records of geothermal tremor and volcanic tremor are really similar (Figure 11, A and C). At the same time, the spectral characteristics of these seismic signals (Figure 11, B and D) are different. The geothermal tremor is characterized by a spectral peak between 3 and 4 Hz while the volcanic tremor is characterized by the peaks at lower frequencies (0.7 and 1.2 Hz).

It was selected the seismic record of a micro-earthquake from the swarm of these events preceding the 5 June 2005 explosion at Volcán de Colima which waveform was similar to the distinct seismic event recorded at Uzon (Figure 12, A and C). Again the volcanic event is characterized by lower peak frequencies (Figure 12 C; between 1.2 and 2 Hz) comparing with the peak frequencies of geothermal event (Figure 12B, between 3 and 8 Hz).

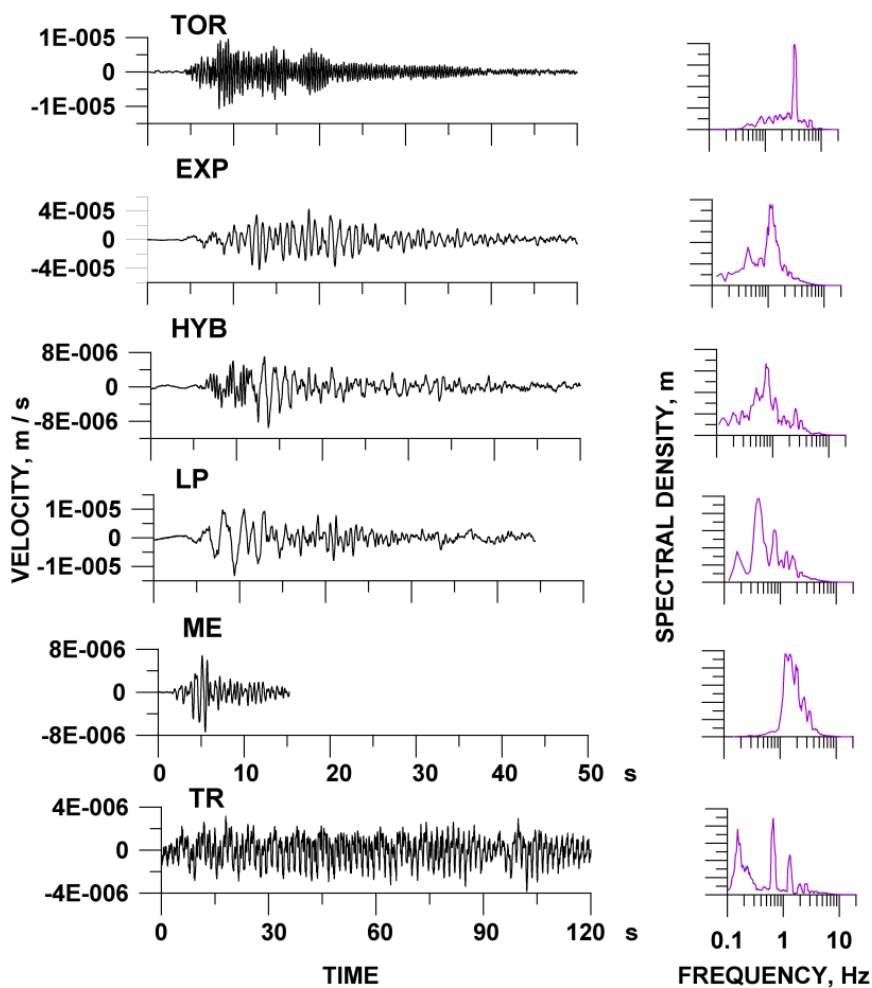


Figure 10. Variety of seismic signals (velocity, vertical component) recorded during the 1998-2011 eruptions at andesitic Volcán de Colima (left side) and corresponding Fourier spectra (right side). TOR, tornillos; EXP, seismic signals associated with magmatic explosive events; HYB, hybrid signals associated mainly with white-cloud degassing; LP, long-period signal recorded at different stages of unrest; ME, micro-earthquake signals, forming together with short hybrid signals the micro-earthquake swarms, associated with extrusive and explosive activity; TR, volcanic tremor recorded at different stages of unrest. All seismic signals were recorded by broadband seismic station at a distance of 4 km from the crater and corrected for instrument response. Courtesy of Colima Volcano Observatory.

So it is possible to say about the similarity of the waveforms of volcanic and geothermal seismic signals but the difference in the spectral contents. This difference in the spectral content indicates the difference in the size of the processes generating the similar seismic waveforms, significantly smaller for geothermal activity.

The geothermal systems, situated within the crater zone of the active volcano, may produce the seismic signals similar also to tornillos (TOR, Figure 10) as was described for the active cone La Fossa of Vulcano volcano, Aeolian islands (Montalto, 1994; Milluzzo et al., 2010).

Tornillos are the sinusoidal waveforms with screw-like envelope profile that may continue for several minutes and are characterized by a few narrow spectral peaks. They have the properties of volcanic tremor and long-period seismic events; therefore, they are not strange for geothermal systems.

Exploitation of the high-temperature geothermal systems may be a reason for occurrences of induced earthquakes which are the product of re-activation of existing local faults. The maximum magnitude of these induced seismic events may reach up to the 4-5 range (Majer et al., 2007).

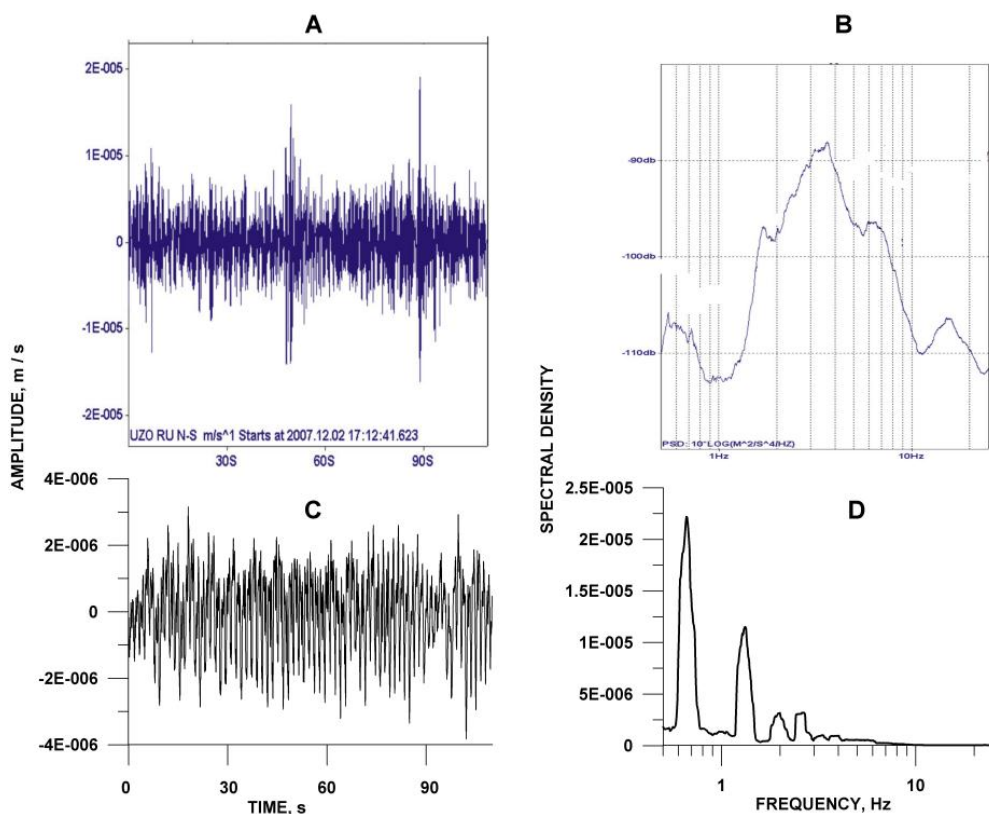


Figure 11. Comparison of the seismic records and corresponding spectra of the geothermal tremor recorded at Uzon geothermal system (A and B) and volcanic tremor recorded at Volcán de Colima (C and D). The short-period seismic station at Uzon was installed at a distance of about 200 m to the north of the boiling mudpot and hot spring; broadband seismic station at Volcán de Colima was installed at a distance of 4 km from the crater. The seismic data of Uzon are taken from (Kugaenko et al., 2009); the seismic data of Volcán de Colima are courtesy of Colima Volcano Observatory.

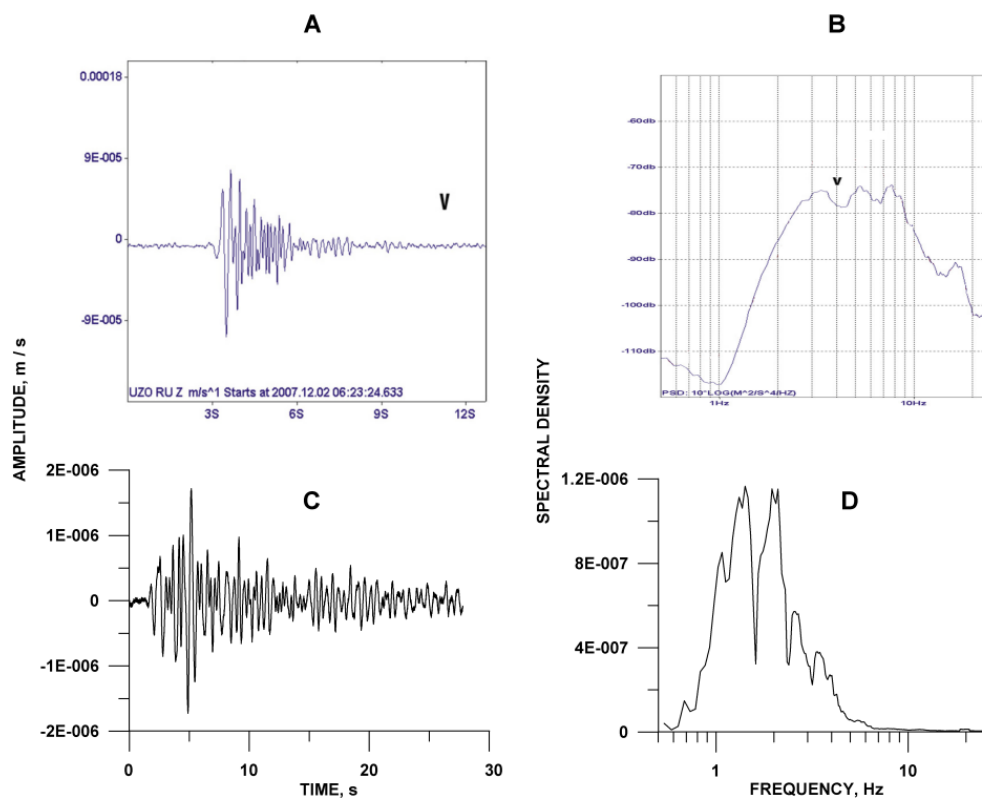


Figure 12. Comparison of the seismic records and corresponding spectra of the geothermal distinct event recorded at Uzon geothermal system (A and B) and the micro-earthquakes (pulga) recorded at Volcán de Colima (C and D). The short-period seismic station at Uzon was installed at a distance of about 200 m to the north of the boiling mudpot and hot spring; broadband seismic station at Volcán de Colima was installed at a distance of 4 km from the crater. The seismic data of Uzon are taken from (Kugaenko et al., 2009); the seismic data of Volcán de Colima are courtesy of Colima Volcano Observatory.

A good example of influence of the geothermal system enhance on the local seismicity may be seen from the case of the Pauzhetkaya geothermal system situated in the Southern Kamchatka, Russia within the East Kamchatka Volcanic Belt, near active volcanoes Kambalny and Koshelevsky (Figure 13). The first exploration wells were drilled in 1957, and the geothermal power station began to operate with a capacity of 7 MW in the end of 1966 (Kiryukhin, 1984). The withdrawal rate gradually increased from 1966 to 1974 (Figure 13A).

The seismic station, installed near the plant in 1959, initially recorded no seismic events within the distance of 20 km from the plant, but beginning from 1973 the swarms of very small local events with magnitude about 1-2 began to occur. The magnitude of seismic events increased during 1975-1977.

Two largest normal-fault events occurred on 17 January 1975 (*mb* 4.6) and on 14 August 1976 (*mb* 4.8) with epicenters situated at a distance of 5-7 km from the plant (Figure 13B). They were felt in the plant building with the modified Mercalli scale intensity V-VI. The seismic activity practically ceased during 1978-1990. New increase in the number of small seismic events was recorded during 1998-2000 (Levina et al., 1980; Chebrov and Kugaenko, 2005).

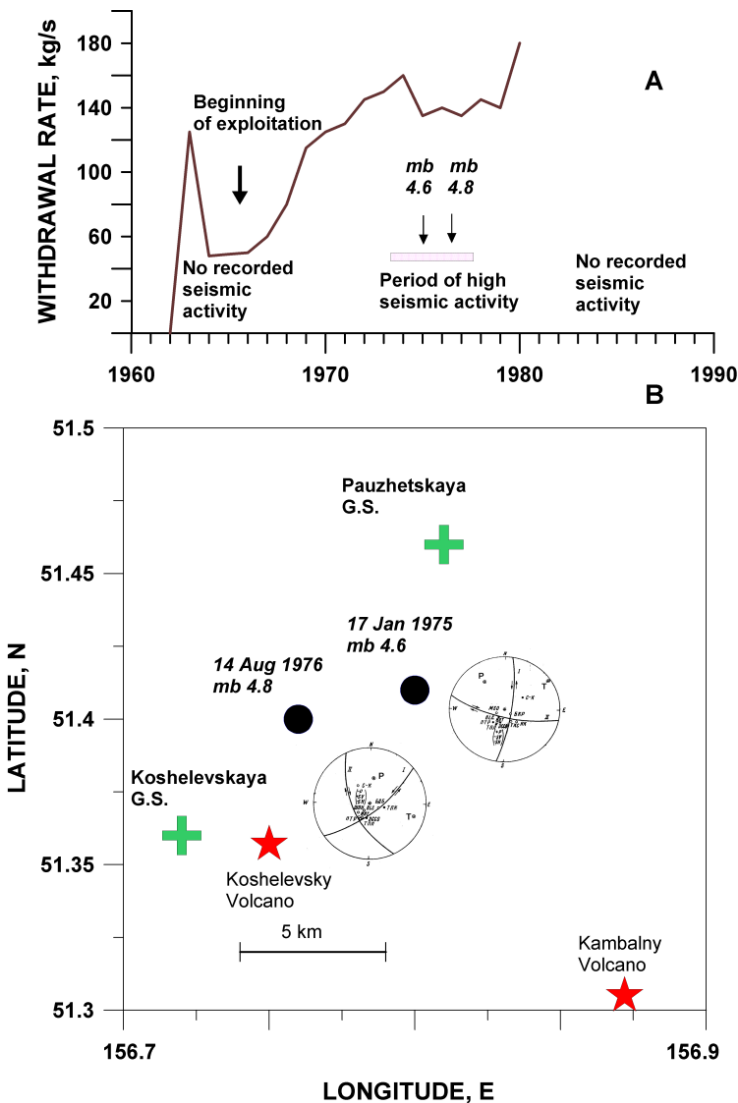


Figure 13. Induced seismicity at the Pauzhetskaya geothermal system, Kamchatka, as a result of its exploitation. A, withdrawal rate at Pauzhetskaya geothermal system; B, epicenters of two largest earthquakes occurring on 17 January 1975 and 14 August 1976 near Pauzhetskaya geothermal plant (shown by filled circles). The focal mechanism solutions of these two events are shown. The curve of withdrawal rate is taken from (Kiryukhin, 1984). The data about the seismic activity are taken from (Levina et al., 1980). Focal mechanisms are constructed by V. Zobin. Magnitudes *mb* of the earthquakes were estimated by U.S.G.S. Crosses shows the Pauzhetskaya and Koshelevskaya geothermal systems; stars show the Kamalny and Koshelevsky active volcanoes.

CONCLUSION

Geothermal seismicity represents a special class of the natural seismic events. They occur as a continuous geothermal tremor (noise) and/or the distinct short seismic events. Comparing with the seismic events associated with volcanic eruptions, the geothermal seismic events are

of higher-frequency indicating the smaller size sources of seismic vibrations. The geothermal seismic events may be useful for understanding of the nature of geyser eruptions and the regime of boiling hot springs.

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